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REFERENCE

Lane, B. H., 1935, Suggestions to authors of papers submitted for publication by the United States Geological Survey: U. S. Geol. Survey, Washington, D. C. (For sale by Superintendent of Documents, Washington, D. C. 40c.)

[American Journal of Science, Vol. 255, December 1957, P. 671-689]

American Journal of Science

DECEMBER 1957

LATE PLEISTOCENE RADIOCARBON CHRONOLOGY IN IOWA[1]

R. V. RUHE, MEYER RUBIN and W. H. SCHOLTES[2]

ABSTRACT. New radiocarbon dates in Iowa permit a grouping of age values and raise new problems in stratigraphic correlation of late Pleistocene deposits in Iowa and adjacent regions. An older group of ages greater than 29,000 years dates the Iowan substage and pre-Iowan deposits. An old group of ages of 22,900 to 25,100 years dates the Farmdale substage. An intermediate group of ages ranges from 14,000 to 17,000 years. Deposition of Tazewell loess occurred in central Iowa during the period 15,000 to 17,000 years before present. A minimum date of the Tazewell-Cary interstadial is 14,000 years. The Cary maximum in the southern extremity of the Des Moines lobe in Iowa was 14,000 years ago. A young group of ages range from 11,900 to 14,000 years and record Cary glaciation in the southern extremity of the Des Moines lobe. Other parts of the lobe previously classified as Mankato are dated at 11,600 to 11,800 years, and antedate Two Creeks interstadial. A younger group of ages approximating 6,500 to 8,200 years record post-glacial major vegetational and climatic changes in north-central Iowa and subaerial erosion and sedimentation in southwestern Iowa.

INTRODUCTION

A previous article (Ruhe and Scholtes, 1955) described the radiocarbon chronology of the late Pleistocene of Iowa, using the dates from samples determined by W. F. Libby at the University of Chicago radiocarbon laboratory. With the advent of the gas counting method and its associated extended range, additional samples from critical sections in Iowa were selected and submitted to the U. S. Geological Survey Radiocarbon Laboratory. As was anticipated, the discovery of new sections containing suitable dating material, combined with the increased precision of the new C-14 dating method, required that some modification and expansion of the previous chronology be made. It is expected that as more samples are dated from Iowa and from adjacent areas of continental glaciation, the details of the glacial events will become clearer and the problems now presented will be closer to solution. The authors therefore consider this more of a progress report than a final statement.

The chronology is described in terms of the sections from which the dated samples have been taken and have been grouped in sequence from the oldest to the youngest.

OLDER GROUPS

Fayette section.—The Fayette section (fig. 1, J) is a cut along the Chicago, Milwaukee, St. Paul, and Pacific Railroad in the center NE¼ sec. 3, T. 92 N., R. 8 W., Fayette County, Iowa, and is one and a half miles southeast

[1] Joint contribution of Soil Survey Investigations, Soil Conservation Service, U. S. Department of Agriculture, Iowa Agricultural Experiment Station, and U. S. Geological Survey. Journal Paper No. J-3203, Iowa Agricultural Experiment Station, Ames, Iowa, Project No. 1250. Publication authorized by the director, U. S. Geological Survey.

[2] Research Geologist, Soil Survey Investigations, SCS, USDA; Geologist, U. S. Geol. Survey; Professor of Soils, Iowa Agric. Experiment Station., respectively.

of the town of Fayette. The cut occurs on the northern flank of a *paha* that trends NE-SE. The section was reported previously by McGee (1891, p. 488).

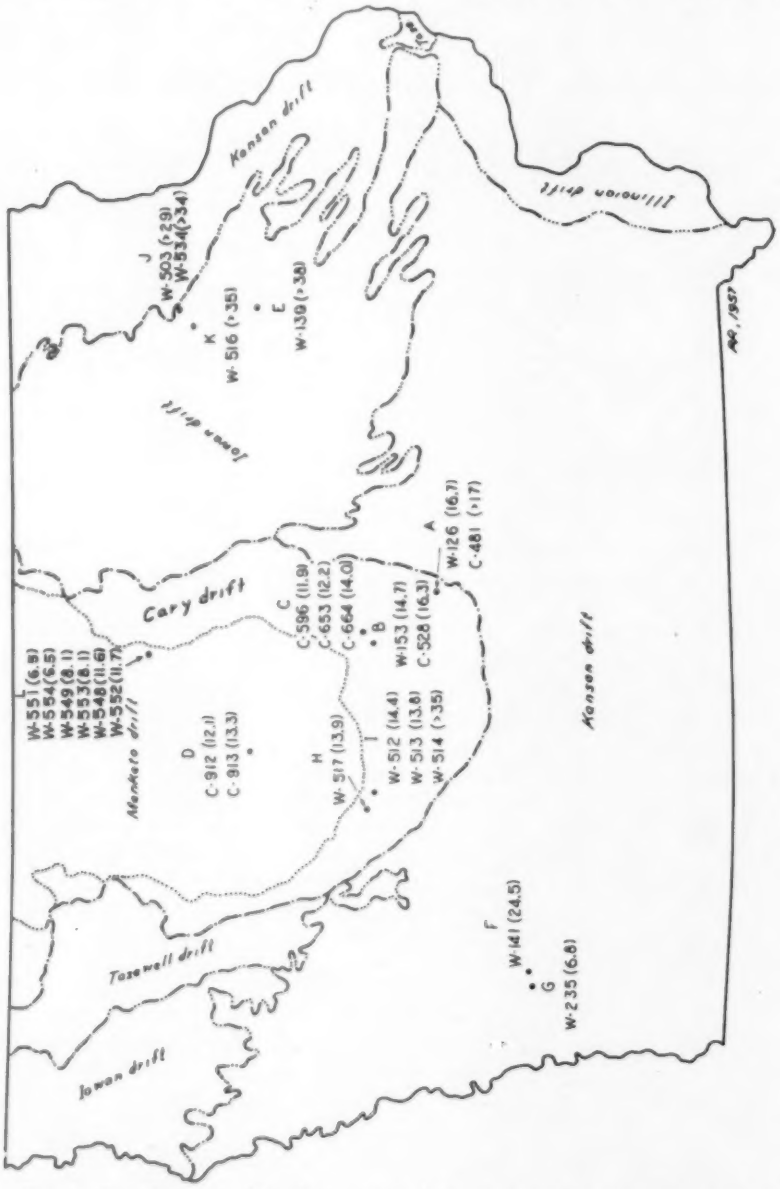

Fig. 1. Locations of radiocarbon sections in Iowa. Location and sample number of radiocarbon section indicated as E W-139 (> 38). Number in parenthesis is age in thousands of years. Glacial-drift borders as shown.

Section description:	Thickness ft.	in.
Iowan till		
4. Yellowish brown (10YR5/4)[3] leached cobbly loam[4] till; mottled with light brownish gray (10YR6/2); friable; subangular blocky; oxidized and leached zone.	9	3
3. Brownish yellow (10YR6/8) to yellowish brown (10YR5/4) calcareous cobbly loam till; firm; coarse angular blocky; oxidized and unleached zone.	5	5
2. Very dark gray (10YR3/1) calcareous cobbly loam till; firm; coarse angular blocky; unoxidized and unleached zone; wood fragments in lower 18 inches with main concentration at contact with underlying zone (radiocarbon sample W-503); zone with wood fragments is McGee's forest bed.	4	10
Pre-Iowan till		
1. Very dark grayish brown (10YR3/2) calcareous clay loam till with dark reddish brown (5YR3/4) iron oxide stain on aggregate faces; firm; medium angular blocky; oxidized and unoxidized and unleached zone transition.	2	7+

A paleosol does not separate stratigraphically the lower till (1) from the basal Iowan till (2). However, the iron oxide stain on the aggregate faces in the lower till (1) suggests that at least an oxidized weathering profile may have been truncated by the overriding Iowan ice.

The Fayette section is located near the margin of the Iowan drift in its type area (fig. 1). Inclusion of abundant wood suggests that Iowan ice encroached on a forested landscape and incorporated fragments of the trees, hemlock[5], in the basal part of the till that subsequently was deposited by Iowan ice. The wood has been dated at $>$ 29,000 years (table 1, W-503), and $>$ 34,000 years in a rerun (W-534).[6]

Maynard section.—The Maynard section (fig. 1, K) is a cut along the Chicago, Rock Island, and Pacific Railroad in the NE¼ sec. 27, T. 92 N., R. 9 W., Fayette County, Iowa, and is a mile south of the town of Maynard. The section was described previously by McGee (1891, p. 489).

Section description	Thickness ft.	in.
Iowan till		
3. Yellowish brown (10YR5/4) leached loam till; friable to firm; angular blocky; oxidized and leached zone.	6	10
2. Brownish yellow (10YR6/8) to yellowish brown (10YR5/4) calcareous loam till; firm; coarse angular blocky; oxidized and unleached zone.	5	0
1. Very dark gray (10YR3/1) calcareous loam till; firm; coarse angular blocky; unoxidized and unleached zone; wood fragments disseminated commonly 30 inches below top of zone and to depth (radiocarbon sample W-516).	5	0+

A lower till, such as in the Fayette section, was not penetrated by shallow borings at the base of the shallow cut.

[3] Munsell soil colors, moist soil.

[4] "Loam" is a texture class which ranges from 7 to 27% clay, 28-50% silt, and $<$ 52% sand.

[5] All woods from Iowa discussed in this paper were identified by Professor D. W. Bensend, Department of Forestry, Iowa State College.

[6] Sample numbers preceded by *W* indicate determination made at the U. S. Geological Survey Laboratory at Washington, D. C. Prefix *C* denotes University of Chicago laboratory determinations.

TABLE 1
Radiocarbon Dates in Iowa

Age group	Substage	Sample	Date	Location	Vegetation
Younger	— epoch	W-551	6,570 ± 200	McCulloch peat bog, Hancock County, Iowa	Grass[1]
		W-554	6,580 ± 200		
		W-235	6,800 ± 300	Middle Silver Creek, Pottawattamie County, Iowa	
		W-549	8,170 ± 200	McCulloch peat bog, Hancock County, Iowa	Oak[1]
		W-553	8,110 ± 200		
Young	Cary substage	W-548	11,660 ± 250	McCulloch peat bog, Hancock County, Iowa	Spruce[1]
		W-552	11,790 ± 250		
		C-596	11,952 ± 500	Cook Quarry, Story County, Iowa	Hemlock
		C-653	12,200 ± 500		
		C-912	12,120 ± 530	Lizard Creek, Webster County, Iowa	Hemlock
		C-913	13,300 ± 900		
Intermediate	Tazewell-Cary interstadial	W-517	13,910 ± 400	Scranton No. 2, Greene County, Iowa	Spruce
		C-664	14,042 ± 1000	Cook Quarry, Story County, Iowa	Hemlock
		W-512	14,470 ± 400	Scranton No. 1, Greene County, Iowa	Fir, hemlock, larch, spruce
		W-513	13,820 ± 400		
	Tazewell substage	W-153	14,700 ± 400	Clear Creek, Story County, Iowa	Hemlock
		C-528	16,367 ± 1000		
		W-126	16,720 ± 500	Mitchellville, Polk County, Iowa	Yew, spruce, hemlock
		C-481	> 17,000		
Old	Farmdale substage	W-141	24,500 ± 800	Hancock, Pottawattamie County, Iowa	Larch
Older	Iowan substage	W-503	> 29,000	Fayette, Fayette County, Iowa	Hemlock
		W-534	> 34,000		
		W-514	> 35,000	Scranton No. 1, Greene County, Iowa	Spruce
		W-516	> 35,000	Maynard, Fayette County, Iowa	Spruce
	Pre-Iowan time	W-139	> 38,000	Independence, Buchanan County, Iowa	Hemlock

[1] Pollen sequence after Lane (1931).

The Maynard section is located in the marginal part of the type Iowan drift, and is approximately six miles southwest of the Fayette cut (fig. 1). At Maynard, as at Fayette, inclusion of wood in the basal part of the Iowan till suggests an overriding of a forested landscape by Iowan ice with resultant incorporation of the wood in the basal part of the Iowan till. The wood has been identified as spruce, and has been dated at > 35,000 (table 1, W-516).

Independence section.—The Independence section (fig. 1, E) is located in a cut along the Illinois Central Railroad in the NW¼ sec. 3, T. 88 N., R. 8 W., Buchanan County, Iowa, approximately 5½ miles east of the town of Independence. The cut is 2½ miles east of Doris station, the type locality for the Iowan drift (Alden and Leighton, 1917, p. 113). The section has been described and discussed previously by Ruhe and Scholtes (1956, p. 264) and by Flint and Rubin (1955, p. 656). However, the section bears more significance than previously accorded now that the Fayette and Maynard sections have been dated.

Section description	Thickness ft.	Thickness in.
Iowan loess		
8. Dark grayish brown (10YR4/2) leached gritty silt; friable; massive; oxidized and leached.	3	2
Iowan till		
7. Pebble band	0	6
6. Yellowish brown (10YR5/8) leached loam till; firm; coarse angular blocky; oxidized and leached zone.	5	8
5. Yellowish brown (10YR5/8) calcareous loam till; firm; coarse angular blocky; oxidized and unleached zone.	2	0
Pro-Iowan (?) or Farmdale (?)		
4. Gray (2.5Y5/0) leached silt; friable; sparse horizontal carbonaceous bands; deoxidized and leached zone.	2	0
Pre-Iowan (?) or pre-Farmdale (?)		
3. Black (2.5Y2/0) leached peat; porous and friable; woody.	1	0
2. Gray (2.5Y5/0) to dark gray (2.5Y4/0) leached silt; friable; massive; deoxidized and leached zone; contains abundant wood fragments; radiocarbon sample (W-139) from this zone, 2 inches below peat.	1	9
1. Gray (10YR5/1) calcareous loam till; firm; coarse angular blocky; unoxidized and unleached zone.	1	6+

In the section (fig. 2) the lower till (1) rises to the surface in the eastern part of the cut where it is truncated by the pebble band. Thus, the continuous pebble band is developed on both the upper till, Iowan (6-5) as well as the lower till, pre-Iowan (1).

The basal, calcareous Iowan till (5) is separated from the buried soil, peat and silt (3-2), by a bed of silt (4) that also is leached of carbonate. The silt (4) grades laterally westward (fig. 2) to oxidized and leached sand, and represents either a pre- or pro-Iowan water-laid sediment. The radiocarbon sample (W-139), dated at > 38,000 years, was contained in the lowermost silt beneath the peat. Wood from the section was identified as hemlock. There is not a well-developed buried soil in the sequence.

The Independence section differs from the Fayette and Maynard sections in that wood of great age is not incorporated in the basal part of the Iowan

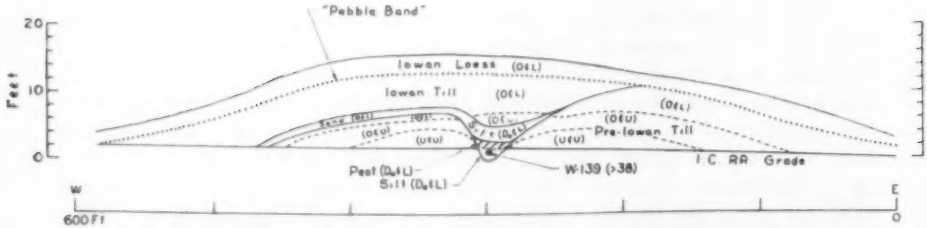

Fig. 2. Scale drawing of Illinois Central Railroad cut, Independence section, Iowa. Symbols: O & L, oxidized and leached; O & U, oxidized and unleached, U & U, unoxidized and unleached; Do & L, deoxidized and leached.

till, but that the wood is separated stratigraphically from the basal Iowan by a buried peat soil and a superposed leached silt and sand bed. The interbedded leached silt and sand does not permit a conclusion that the Iowan ice did or did not advance into the area soon after the peat formed.

Alden and Leighton (1917, p. 95, 99, 100, 111, 202) have described sections in the Iowan drift region in northeastern Iowa where basal Iowan till is underlain by a silt which, in turn, mantles older Pleistocene deposits.

Scranton No. 1 section.—(See Intermediate Group.) In this section (fig. 1, I) an unoxidized and unleached till occurs in the base of the section, and is overlain by an unoxidized and unleached silt that contains a spectacular buried forest in the uppermost part. A paleosol or weathering zone does not separate the lower till from the overlying silt. The silt, in turn, is mantled by unoxidized and unleached Cary till.

The lower part (fig. 4, barren zone) of the buried silt in the section is correlative stratigraphically with the post-Farmdale loess where that loess is the surficial deposit beyond the limits of the Cary drift (figs. 1; 3, F, H, I), and thus may be considered early Wisconsin in age. Because the buried silt in the Scranton section is not separated from the underlying till by a paleosol or weathering zone, the lowermost till also is considered early Wisconsin in age. A wood sample (W-514) was extracted 13½ feet below the top of the lowermost till, and is dated at greater than 35,000 years.

Thus, this section represents the third location (Fayette, Maynard, Scranton No. 1) where wood beyond the present range of radiocarbon dating methods was collected from Iowan or early Wisconsin till. At the fourth location (Independence), an interbedded leached silt and sand separates the Iowan from the horizon of "old" wood.

Hancock section.—The Hancock section (fig. 1, F) is a cut on the Chicago, Rock Island, and Pacific Railroad in Pottawattamie County, Iowa, approximately 2½ miles southwest of the town of Hancock. The section was described in detail by Ruhe and Scholtes (1956, p. 264-265). A wood sample (W-141) was extracted from the A horizon of a buried soil in a loess sheet 3 to 4 feet thick that overlies Loveland loess. Above the sample horizon there is a third loess sheet approximately 30 feet thick. The uppermost loess contains two faunal zones. Based upon Leonard's (1952, p. 11-14) faunal zonations, the uppermost zone is indicative of the Tazewell and the lowermost zone

of the Iowan. On the basis of the stratigraphic position of the radiocarbon horizon beneath an Iowan faunal zone and above the Sangamon paleosol in Loveland loess, the sample horizon was classified by Ruhe (1954, p. 640) as Farmdale. The age of the sample, 24,500 ± 800 years, is in excellent agreement with radiocarbon values (Suess, 1954, p. 470) for the type Farmdale in Illinois at 22,900 ± 900 and 25,100 ± 800 years. Wood from the Hancock section was identified as larch.

Correlation of deposits of Farmdale and Iowan substages.—It is generally recognized that the Farmdale is the earliest substage of the Wisconsin glacial stage and that the Iowan substage was subsequent to the Farmdale (Leighton and Willman, 1950, p. 602; Horberg, 1953, p. 12; Shaffer, 1956, p. 10). Leighton and William defined the Farmdale substage in Illinois as pro-Wisconsin deposits between the Sangamon weathered zone and Iowan loess, and they defined the Iowan substage in Illinois as the loess between the Farmdale and Shelbyville [Tazewell] deposits. Shaffer (1954, p. 455), with qualifications, reconfirmed the younger age of Iowan in eastern Iowa by stating: "If the high terrace of Lake Calvin proves to be equivalent to the Tazewell (Shelbyville) then the Shelbyville portion of the Tazewell and the Iowan would be contemporaneous, and on the basis of priority Shelbyville would be redefined as Iowan. These matters merit further field investigation."

The stratigraphic relationships of Farmdale, Iowan, and Tazewell deposits were determined originally by tracing the loess that occurs on the Iowan drift in eastern Iowa eastward to the margin of the Shelbyville drift in western Illinois (Leighton, 1931, p. 46-51). At the margin of the Shelbyville drift, part of the loess passes under and part over the drift. Paleosols do not separate the lower loess from the interbedded till, or the till from its overlying loess. The lower loess is of Iowan age, and the interbedded till and the upper loess are of Tazewell age (Kay and Leighton, 1933, p. 673). The lower loess (Iowan) overlies a loess that contains a paleosol and youthful profile of weathering in its uppermost part. This basal loess is the "late Sangamon loess" of earlier classification (Leighton, 1931, p. 48), but now the Farmdale loess of current classification (Leighton and Willman, 1950, p. 602-603).

Thus, the classification and correlation of the sequence Farmdale, Iowan, and Tazewell fundamentally is based upon the tracing of the Peorian loess, or more specifically the Iowan increment of the loess, and determining the relationships of the increment to overlying and underlying deposits. Because the increment is not separated from the rest of the Peorian loess by a paleosol, identification and classification of the increment may be somewhat tenuous. For example, in Shaffer's (1956, p. 16-17) recent delineation of Farmdale drift in northwestern Illinois, a loess between the Farmdale and tills of the Shelbyville sub-lobe is classified in all described sections as *Iowan or pro-Shelbyville [pro-Tazewell] loess.*

Radiocarbon dates from the type Iowan drift in Iowa and Farmdale loess in Iowa and the type Farmdale in Illinois suggest that the Iowan may be older than the Farmdale. The ages of wood in unoxidized and unleached Iowan till range from more than 29,000 years to more than 35,000 years (Fayette, W-503, > 29,000 years, W-534, > 34,000 years; Maynard, W-516, >

35,000 years; Scranton No. 1, W-514, > 35,000 years). The ages of wood from the Farmdale range from 22,900 years to 25,100 years (Hancock, Iowa, W-141, 24,500 ± 800 years; Farm Creek, Illinois, W-68, 22,900 ± 900 years and W-69, 25,000 ± 800 years).

In contrast to the "greater than" dates derived from wood in the Iowan till of Iowa, determinations made on wood found between Farmdale loess and overlying Tazewell till in the supposed loess correlative in Illinois, have yielded dates of from 20,340 to 22,450 years. This loess has been dated by three samples (Farm Creek, Illinois, W-349, 20,340 ± 750; Rock Creek, Illinois, W-483, 20,500 ± 600; East Bureau Creek, Illinois, W-334, 22,450 ± 1,000) and the dates obtained are in complete accord with the dates from wood in the overlying and underlying deposits. The fact that these dates are what one might expect for a loess between Farmdale and Tazewell deposits does not verify the correlation of the loess as Iowan. Rather, the opposing dates of the Iowan till of Iowa, and the Iowan loess of Illinois present an unresolved problem of field correlation.

Identification of an Iowan faunal zone above the Farmdale at Hancock by Ruhe (Rubin and Suess, 1955, p. 484) was based upon the faunal zonations of Leonard (1952, p. 11-14). Leonard's faunal zonations, in turn, are based upon the somewhat tenuous physical stratigraphic correlations of Farmdale, Iowan, and Tazewell in eastern Iowa and western Illinois (Leonard, 1952, p. 15; Frye and Leonard, 1952, p. 128).

Mitigative evidence in the lack of agreement between radiocarbon ages and accepted stratigraphy of Farmdale and Iowan may occur in the Independence section in Iowa (fig. 2). There the basal calcareous Iowan till is separated from the buried peat (<38,000 years) by an interbedded, leached silt and sand. Is this leached zone correlative of Farmdale or of the pro- or earliest Iowan basal, leached loess (fig. 3, E-F)? The problem warrants further investigation.

However, it should be noted that three sections of Iowan age in Iowa contain wood of radiocarbon age similar to the old till in Ohio, typified by the Sidney section (Flint and Rubin, 1955, p. 651; Forsyth and La Rocque, 1956; Rubin, 1956; Goldthwait, in press), that may indicate a pre-Farmdale Wisconsin glaciation. If this be substantiated, then the name Iowan would have priority for this earliest of post-Illinoian glaciations.

INTERMEDIATE GROUP

In Iowa a considerable number of radiocarbon dates are included in an intermediate group that is divisible into two age ranges, one of 15,000 to 17,000 years, and a second at 14,000 years. Sections of the older category have been described previously: Mitchellville section (fig. 1, A; Ruhe and Scholtes, 1955, p. 83-84) and Clear Creek section (fig. 1, B; Ruhe and Scholtes, 1955, p. 84-88). These two sections are very similar stratigraphically. A thin Cary till caps a buried loess within which, and at a considerable number of feet below the till, forests beds occur. The Mitchellville forest bed has been twice dated (table 1: C-481, >17,000 years; W-126, 16,720 ± 500 years)

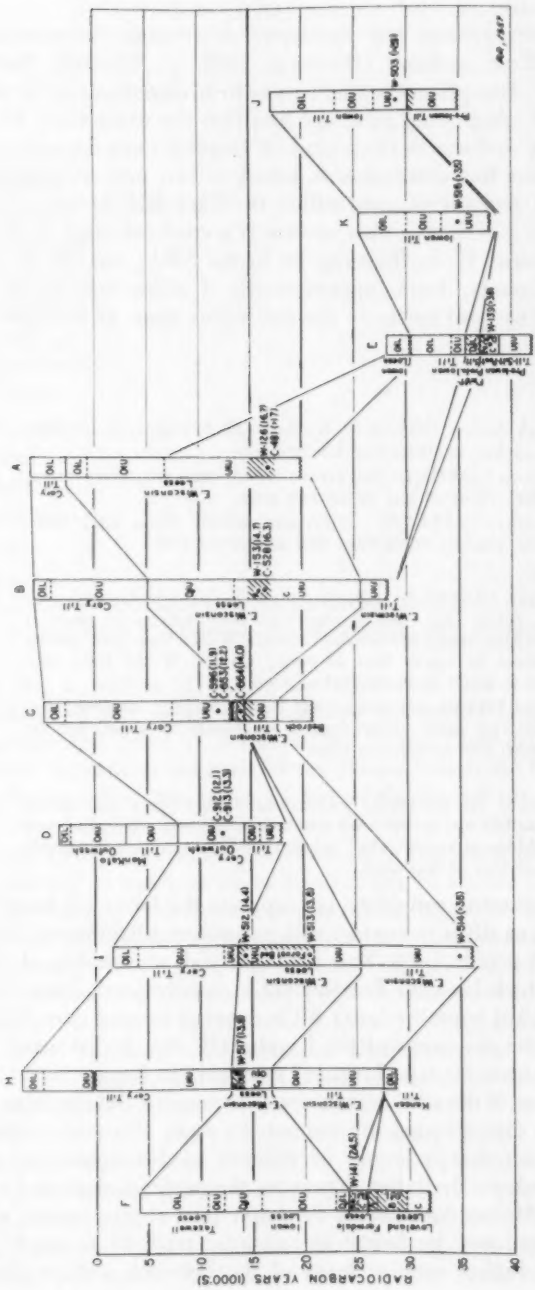

Fig. 3. Correlation of radiocarbon sections in Iowa. Symbols: See fig. 2. A, B, C are horizons of buried soils. W-139 (> 38) is sample location and number. Number in parenthesis is age in thousands of years.

as has the Clear Creek forest bed (C-528, 16,367 ± 1,000 years; W-153, 14,700 ± 400 years).

Considerable argument has developed concerning the correlation and significance of these sections (Horberg, 1955, p. 280-281; Ruhe, 1956; Leighton, 1956). The present writers reconfirm classification of the buried loess as Tazewell rather than pro-Cary and that the dates from 15,000 to at least 17,000 years conform to the period of Tazewell loess deposition. Further evidence to support this conclusion is found in two new sections (Scranton No. 1 and No. 2) recently exposed within the Cary drift in central Iowa.

Scranton No. 1 section.—This section is a road cut (figs. 1, I; 4) along the recently relocated U. S. Highway 30 in the SW¼ sec. 33, T. 84 N., R. 31 W., Greene County, Iowa, approximately 4 miles east of the town of Scranton, Iowa. The road cut is on the east valley slope of Raccoon River.

Section description	Thickness ft.	Thickness in.
Cary till		
6. Brown to dark brown (10YR4/3) leached cobbly loam till; friable; subangular blocky; oxidized and leached zone.	3	6
5. Yellowish brown (10YR5/6) calcareous cobbly loam till; firm; coarse angular blocky; oxidized and unleached zone.	12	4
4. Very dark gray (2.5YR3/0) calcareous cobbly loam till; firm; coarse angular blocky; unoxidized and unleached zone.	2	0
Early Wisconsin silt		
3. Very dark gray (2.5YR3/0) calcareous silt loam; friable; massive; fossiliferous; forest bed—tree stumps with root systems in place at many levels within zone; radiocarbon sample W-512 from root system of tree in place in upper foot of zone; sample W-513 from root system of tree in place in lower foot of zone.	11	0
2. Dark gray (2.5YR4/0) calcareous silt loam; friable; massive; fossiliferous; "barren zone" disseminated fragments of wood, but no tree stumps with root systems in place.	6	8
Early Wisconsin till		
1. Greenish yellow (5GY5/1-6/1) calcareous cobbly loam till; firm; coarse angular blocky; unoxidized and unleached zone; polished and varnished pebbles at top of till, "pebble band"; radiocarbon sample W-514 in basal foot of this zone.	13	6+

A paleosol or weathering zone does not separate the lower till from the overlying silt. Calcareous silt is in contact with calcareous till, although interrupted by an intercalated pebble band. This is the normal relationship of thick loess on Iowan till or thick loess on Tazewell till in northwestern Iowa. The radiocarbon sample W-514 from the lower till is dated at greater than 35,000 years and conforms to the age range of the Fayette (W-503, W-534) and Maynard (W-516) samples from the type Iowan of northeastern Iowa.

The lower part of the silt (barren zone) is massive and contains a gastropod fauna that is disseminated throughout the zone. Physical characteristics of the silt are such that it would be difficult to distinguish the silt from recognized loess where the latter occurs as the surficial deposit beyond the limits of the Des Moines lobe. Thus, the lower part (barren zone) of the silt is considered to be loess. Its faunal assemblage (table 2) is nearly identical to the uppermost faunal zone of loess of the Hancock section (fig. 1, F),

TABLE 2
Faunas of Radiocarbon Sections

Species	Hancock						Scranton No. 2			Scranton No. 1			Clear Creek					
	Lower zone			Upper zone									Below Forest Bed			Above Forest Bed		
	I	Tr	T	I	Tr	T	I	Tr	T	I	Tr	T	I	Tr	T	I	Tr	T
Columella alticola						*			*			*						*
Euconulus fulvus									*			*						*
Hendersonia occulta						*			*									*
Succinea ovalis						*			*			*						*
Discus cronkhitei		*	*					*	*		*	*		*	*		*	*
Discus shimeki		*	*		*	*					*	*						
Succinea grosvenori		*	*		*	*		*	*		*	*		*	*		*	*
Vertigo modesta		*	*					*	*		*	*					*	*
Hawaii minuscula	*	*	*				*	*	*				*	*	*			
Pupilla blandi							*	*	*	*	*	*	*	*	*	*	*	*
Pupilla muscorum							*	*	*				*	*	*			
Vallonia gracilicosta	*	*	*	*	*	*	*	*	*	*	*	*	*	*	*	*	*	*
Succinea avara							*	*		*	*		*	*		*	*	
Lymnaea parva										*			*					

I—Iowan, Tr—transition, T—Tazewell (Faunal zonation after Leonard, 1952).

Scranton No. 1 section (H), and the uppermost faunal zone of the Clear Creek section (B). It appears certain that the buried loess of the Scranton No. 1 section and those of Scranton No. 2 and Clear Creek are the same stratigraphic horizon as the loess of the Hancock section (fig. 3, F-H-I-B).

The forest bed in the upper 11 feet of the buried silt is an outstanding feature. Tree stumps with root systems in place occur at many levels throughout the zone (pl. 1, A). A horizontal bench in the forest bed cut during construction showed as many as 15 to 20 trees per 50 square feet of area. Wood from the section has been identified as hemlock, larch, fir, and spruce.

However, the silt containing the forest bed is separated from the underlying "barren zone" by an erosion surface. Fragments of wood and gastropod shells are concentrated along a bedding plane, almost continuous throughout the section, at the base of the forest bed. Other bedding planes with concentrations of wood fragments and gastropod shells occur within the forest bed. These evidences suggest that the silt which contains the forest bed is a water-laid sediment that was derived from a source of early Wisconsin loess.

Wood in the forest bed just below the contact of the Cary till is dated at 14,470 ± 400 years (table 1, W-512). Trunks of trees rooted in this zone were bent over toward the south. The date of 14,470 ± 400 years is in excellent agreement with a date (table 1: C-664, 14,042 ± 1,000 years) from the Cook Quarry section (Ruhe and Scholtes, 1955, p. 88-90) where wood was collected just below the base of the Cary till. These radiocarbon values

PLATE 1

Spruce tree rooted in place in forest bed of Scranton No. 1 section, Iowa. Trunk of a second tree is in right background.

Buried soil in early Wisconsin loess between overlying Cary till and underlying early Wisconsin till in Scranton No. 2 section, Iowa. A_1, C_1, C_2 are horizons of buried soil.

date the earliest advance of Des Moines lobe Cary ice in the latitude of the sections (fig. 1).

Wood collected 10 feet below sample W-512 but just above the basal bedding plane of the forest bed silt is dated at 13,820 ± 400 years (table 1, W-513).

In the Scranton No. 1 section the buried forest is a gigantic Regosolic paleosol[7] that stratigraphically separates the loess (barren zone) from the overlying Cary till. The loess is most logically Tazewell in age and not pro-Cary as Horberg (1955, p. 281) has suggested for the similar stratigraphic sections at Clear Creek and Mitchellville (fig. 1, B and A).

Scranton No. 2 section.—This section (figs. 1, H; 4) is a cut on country road M on the south valley slope of Raccoon River in the NE¼ sec. 26, T. 84 N., R. 32 W., Greene County, Iowa, approximately 2½ miles north of the town of Scranton, Iowa.

[7] A *Regosol* is a soil developed in unconsolidated material, and which has few or no clearly expressed soil characteristics. A Regosol generally has an A-C profile. A *paleosol* is a "fossil" soil that was formed on a landscape during the geologic past. Sedimentation subsequent to pedogenesis buried and preserved the soil profile. This kind of paleosol is a buried soil.

Section description	Thickness ft.	Thickness in.
Cary till		
14. Dark brown (10YR3/3) leached cobbly loam till; firm; coarse angular blocky; unoxidized and leached zone.	3	0
13. Yellowish brown (10YR5/4) and dark yellowish brown (10YR4/4) mottled with gray (10YR5/1) calcareous cobbly loam till; firm; coarse angular blocky; oxidized and unleached zone.	28	0
12. Very dark gray (10YR3/1) calcareous cobbly loam till; firm; coarse angular blocky; unoxidized and unleached zone.	12	10
Early Wisconsin loess		
11. Very dark gray (10YR3/1) to very dark grayish brown (10YR3/2) weakly calcareous silt loam; friable; weak fine granular; A_{11b}[8] horizon of Tazewell-Cary interstadial soil.	0	6
10. Very dark grayish brown (10YR3/2) to dark brown (10YR3/3) very weakly calcareous silt loam; friable; weak fine granular; A_{12b} horizon of Tazewell-Cary interstadial soil.	0	6
9. Dark grayish brown (2.5YR4/2) to very dark grayish brown (2.5YR3/2) leached silt loam; friable; massive; C_{1b} horizon of Tazewell-Cary interstadial soil; tree stumps with root systems in place in A_1-C_1 horizons of buried soil (radiocarbon sample W-517).	0	7
8. Dark grayish brown (2.5YR4/2) weakly calcareous silt loam; friable; massive; fossiliferous; C_{2b} horizon of Tazewell-Cary interstadial soil.	1	8
7. Olive brown (2.5YR4/4) calcareous silt loam; friable; massive; fossiliferous, oxidized and unleached loess.	0	9
Early Wisconsin till		
6. Dark yellowish brown (10YR4/4) calcareous cobbly loam till with reddish brown (2.5YR4/4) iron oxide stain on aggregate faces in upper foot of zone; firm; coarse angular blocky; oxidized and unleached zone.	6	11
5. Gray (10YR5/1) to grayish brown (10YR5/2) calcareous cobbly loam till; firm; coarse angular blocky; unoxidized and unleached zone.	9	2
Kansan till		
4. Yellowish brown (10YR5/4) leached cobbly clay loam[9] till; firm; plastic; coarse angular blocky; oxidized and leached zone.	1	10
Pro-Kansan sand and gravel		
3. Strong brown (7.5YR5/6) leached gravelly sand; bedded.	2	6
2. Yellowish brown (10YR5/8) leached sand.	1	6
1. Strong brown (7.5YR5/6) leached sand; iron-oxide cemented, indurated pan (must be broken with hammer); underlain by yellowish brown, leached, friable, bedded sands and gravels.	1	0

The lower till and underlying sands and gravels (fig. 4) are probably pre-Wisconsin (Kansan) in age. The horizons are leached to a combined depth of more than 10 feet. The till above the Kansan is considered early Wisconsin in age because of its relationship to the overlying buried loess. Oxidized and unleached till is in contact with oxidized and unleached loess. A paleosol or weathering zone does not separate the two. This conformable relationship is characteristic where thick loess overlies Iowan and Tazewell tills in northwestern Iowa.

[8] Soil horizon nomenclature follows Soil Survey Manual (1951, p. 173-188).

[9] "Clay loam" is a texture class which ranges from 27 to 40% clay, 20 to 45% sand, and the remainder of silt.

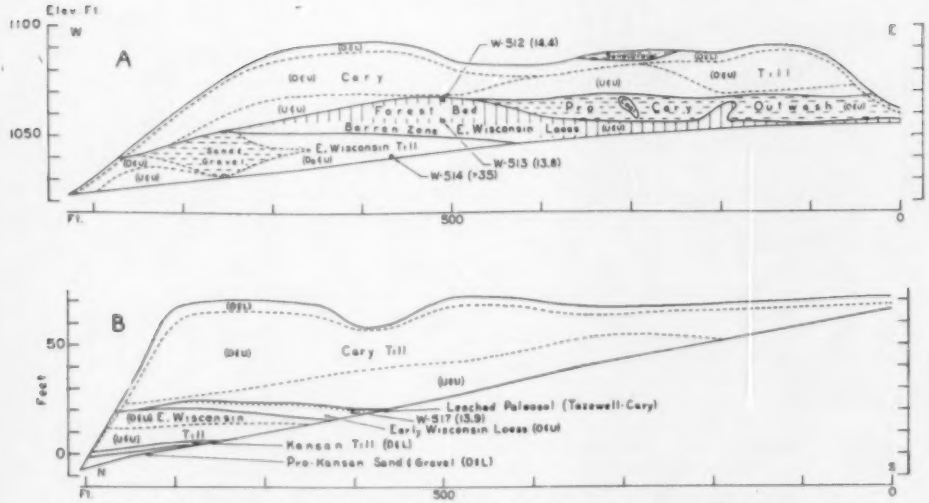

Fig. 4. Scale drawings of Scranton No. 1 section, Iowa (A) and Scranton No. 2 section, Iowa (B). Symbols: see fig. 2.

The buried loess in the section contains a fauna that is almost identical to the faunas of the surficial Tazewell loess in the Hancock section, the buried loess in the Scranton No. 1 section, and the upper zone of the buried loess in the Clear Creek section (table 2). This is one of the bases on which the buried loess of the Scranton No. 2 section is classified as early Wisconsin. The loesses, surficial and buried, of the four sections undoubtedly are correlative (fig. 3, F-H-I-B).

A Regosolic paleosol (pl. 1, B) occurs in the uppermost part of the buried loess which is overlain by Cary till. The paleosol is weakly developed. There is no clay accumulation in the profile (fig. 5). Organic matter has accumulated in greatest quantity in the upper 12 inches of the paleosol. Distribution of organic matter with depth in the profile is comparable to modern Regosols (fig. 5). The paleosol was partially leached of carbonates[10] to a depth of 19 inches (fig. 5), although resaturation by carbonates from the overlying calcareous Cary till increases progressively upward toward the uppermost horizon of the buried loessial soil. The interbedded loess is differentiated distinctly from the overlying and underlying tills on the basis of sand content (fig. 5). Similarly, organic matter and carbonate contents differentiate the loess from the tills (fig. 5), and emphasize the paleosol development.

Wood from a tree stump rooted in place in the A_1-C_1 horizons of the buried soil has been dated at 13,910 ± 400 years (table 1, W-517). The wood was identified as spruce. The date is in excellent agreement with dates from Scranton No. 1 section at 14,470 ± 400 years and Cook Quarry section at 14,042 ± 1,000 years for comparable stratigraphic horizons. These dates

[10] In the field the horizon of the paleosol from 12 to 19 inches did not effervesce in strong concentrations of acid. The horizon was recorded as leached.

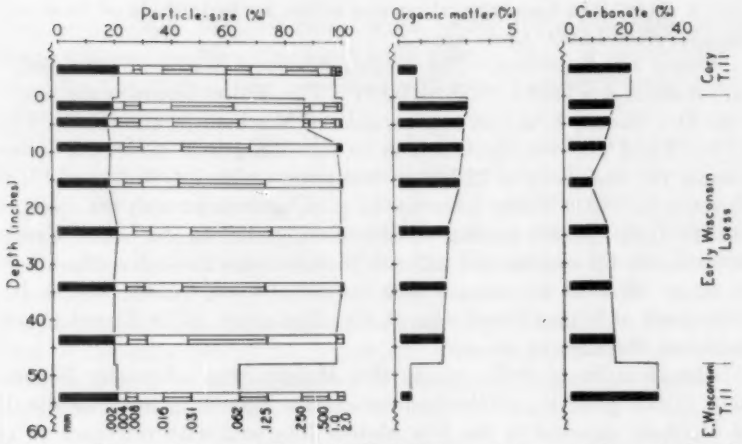

Fig. 5. Characteristics of paleosol in Early Wisconsin loess in Scranton No. 2 section, Iowa.

place the maximum of Cary ice advance of the Des Moines lobe at this latitude (fig. 1, H, I, C) as approximately 14,000 years before present. The maximum limit of Cary ice advance, to the city of Des Moines, is only 30 miles south of the latitude of the three sections, so the Cary maximum in Iowa can be dated at approximately 14,000 years ago.

The occurence and dating of the tree in a paleosol beneath Cary till provides a minimum date of 14,000 years for the Tazewell-Cary interstadial in Iowa. The occurrence of the paleosol in the uppermost part of the loess also lends support to the conclusion that the loess buried beneath Cary drift in the Des Moines lobe area is Tazewell in age.

It is shown therefore that deposition of Tazewell loess was occurring in central Iowa during the period from 15,000 to at least 17,000 years ago (table 1: Clear Creek at 14,700 $\pm$ 480, 16,367 $\pm$ 1,000; Mitchellville at 16,720 $\pm$ 500, $>$ 17,000), but that the deposition terminated prior to 14,000 years ago.

YOUNG GROUP

Two sections in central Iowa have radiocarbon dates that range from 11,900 to 13,300 years. These dates apply to the drifts of the Des Moines lobe.

Cook Quarry section.—This section has been described previously (fig. 1, C; Ruhe and Scholtes, 1955, p. 88-90). Wood extracted from the unoxidized and unleached Cary till has been dated at 11,952 $\pm$ 500 years (table 1, C-596) and 12,200 $\pm$ 500 years (C-653). The section occurs in the marginal part of the Des Moines lobe that was reclassified as Cary in age (Ruhe, 1952, p. 47-49).

A comparison of these dates with the radiocarbon date from an horizon just below the base of the Cary till in the same section (table 1, C-664) and with the dates of the sections Scranton No. 1 and No. 2 (table 1, W-512,

W-517) shows that Cary glaciation was active in the latitude of these sections from 11,900 to 14,000 years ago.

Lizard Creek section.—The section has been described previously (fig. 1, D; Ruhe and Scholtes, 1955, p. 90-91). The section occurs in the inner part of the Des Moines lobe that was considered Mankato in age (Ruhe, 1952, p. 52-54). Wood that was contained in an outwash gravel between a lower and an upper till, was dated at 12,120 ± 530 years (table 1, C-912) and 13,300 ± 900 years (C-913). These dates are in good agreement with the young dates from the Cook Quarry section (C-596, C-653). But in the Lizard Creek section an upper till overlies the outwash that contains the radiocarbon horizon. The upper till must be younger than the till of Cook Quarry (fig. 3, B) and the outwash of Lizard Creek (fig. 3, C). The upper till in Lizard Creek was considered Mankato in age.

Classification of drifts of the Des Moines lobe.—Recently Wright and Rubin (1956, p. 625), on the basis of certain difficulties in delineating Cary and Mankato deposits in the Des Moines lobe and with reference to radiocarbon ages that antedate the Two Creeks interstadial, suggested that the whole of the Des Moines lobe should be placed in the Cary. Zumberge and Wright (1956, p. 65-81) have reaffirmed in greater detail this classification. The arguments of reassigning the whole of the Des Moines lobe to Cary are based primarily on the evidence that most of the radiocarbon dates of the Des Moines lobe are older than Two Creeks (11,400 years).

It should be noted, however, in the extreme southern portion of the Des Moines lobe that Cary events are recorded by radiocarbon dates that range from a maximum of 14,000 years to a minimum of 11,900 years. The latter age is approaching Two Creeks time. All of these events are recorded in front of the Altamont moraine, which was recognized by Ruhe (1952, p. 52-54) as the margin of the Mankato of the Des Moines lobe.

Further evidence, that portions other than the outermost margin of the Des Moines lobe in Iowa may antedate Two Creeks, is found in the McCulloch peat bog which is located in the center S½ sec. 32, T. 94 N., R. 24 W., Hancock County, Iowa (fig. 1, L). In the bog 3½ feet of black (10YR5/1) to very dark brown (10YR2/2) fibrous, woody peat overlies a gray (10YR5/1), calcareous, fossiliferous, organic muck, Six inches of gray sand and gravel occur at the base of the bog sediments that overlie gray, calcareous till. The top of the till is 14 to 15 feet below the surface of the bog. The bog occurs at the inner margin of the Altamont moraine, and occurs also 16½ miles southeast of the outer margin of the massive Algona moraine.

Lane (1931) has described in detail the pollen sequence of the McCulloch peat bog.

Two radiocarbon dates have been determined from the basal 12 inches of organic muck of the bog sediment. An age of 11,660 ± 250 years (table 1, W-548) was determined from the fraction of the muck insoluble in hot NaOH, and a date of 11,790 ± 250 years (W-552) from the fraction of the sample from the same horizon that was soluble in hot NaOH. These separations of the lignin from the cellulose were made in the laboratory to eliminate the possibility of contamination by younger organic matter from above, the lignin

being the most mobile. Both of these values, which are a minimum for the underlying till, predate Two Creeks (11,400 years). Thus, it appears that a portion of the Des Moines lobe associated with the Altamont moraine also is older than Two Creeks.

But if the whole of the Des Moines lobe is to be considered Cary (i.e., pre-Two Creeks), many more events must be crowded in the period 11,700-11,600 to 11,400 years. Construction of the massive Algona moraine in Iowa is required. Then too, the many morainal systems of the lobe in Minnesota must be considered. Before the whole of the Des Moines lobe can be classified as Cary, more evidence must be available than is presently at hand.

Leighton's (1957, p. 109) suggestion may be pertinent, that the Valders be added to the substages of the Wisconsin, succeeding the Mankato. Thus, the Mankato would be maintained as a separate substage but of pre-Two Creeks age rather than post-Two Creeks as presumed previously.

YOUNGER GROUP

Middle Silver Creek section.—This section (fig. 1, G) is a gully fill exposed in a cut along the relocated Chicago, Rock Island, and Pacific Railroad, along the south line sec. 13, T. 76 N., R. 41 W., Pottawattamie County, Iowa. The gully fill truncates progressively downward a late Wisconsin gleyed zone in the upper part of post-Farmdale loess (Ruhe and Scholtes, 1956, p. 271-272), post-Farmdale loess, Farmdale loess, and Sangamon soil in Loveland loess. Thus, the gully fill must post-date late Wisconsin. Carbon extracted by combustion from the basal foot of the fill of 17 feet is dated at 6,800 ± 300 years (Rubin and Suess, 1956, p. 446, W-235).

The gully fill of Middle Silver Creek section is part of a cut and fill surface that occurs in regional distribution throughout southwestern Iowa, and records post-glacial or Recent subaerial erosional modification of the landscape.

McCulloch peat bog section.—Lane (1931, p. 167) on the basis of pollen analyses showed a sequence of changes in the peat bed from the base upward which are: (1) spruce forest, (2) mixed fir, birch, and spruce, (3) birch with fir and oak, (4) oak with birch and grasses, and (5) grassland vegetation with two intercalated strong semi-arid elements indicated by major proportions of amaranths.

Radiocarbon samples were collected from Lane's spruce zone, coniferous-deciduous forest transition zone, and the zone indicating the beginning of grass dominance. The basal spruce zone is dated at 11,660 ± 250 years (fig. 1L; table 1, W-548) and 11,790 ± 250 years (W-552). The coniferous-deciduous forest transition is dated at 8,170 ± 200 years (W-549) and 8,110 ± 200 years (W-553). The beginning of the dominance of grass is dated at 6,570 ± 200 years (W-551) and 6,580 ± 200 years (W-554).[11]

Lane (1931, p. 169) inferred from the pollen sequence a climatic change from cool, moist of the coniferous period to a warming trend which accompanied the change from coniferous to deciduous forms (now dated at

[11] The first of each pair of values was determined from the fraction insoluble in hot NaOH, and the second of each pair from the fraction soluble in hot NaOH.

11,600 to 8,100 years), and a gradual dessication of the climate culminating just prior to grass dominance (now dated at 8,100 to 6,500 years). The warm, dry grassland climate has been continuous from 6,500 years to the present, with the exception of a stronger semi-arid element during that time.

Lane's "culmination of climatic dessication just prior to grass dominance" encompasses the age of the basal gully fill in loess in southwestern Iowa, which is dated at 6,800 ± 300 years (table 1, W-235). Such climatic dessication may have prepared the loessial landscape for severe gullying.

SUMMARY

The following tentative conclusions seem valid from the evidence at hand. It is recognized that further field and laboratory studies are necessary to test their validity.

1. A possible interpretation of radiocarbon dates from Iowa places the Iowan substage in an older position than the Farmdale, rather than younger, as hitherto believed. Farmdale is dated in Iowa at 24,500 years B.P. (before present); Iowan till yields wood with dates beyond the present range of radiocarbon dating.

2. Tazewell loess was being deposited in Iowa from sometime prior to 17,000 to approximately 15,000 years ago.

3. The Cary advance of the Des Moines lobe reached its maximum southern extent approximately 14,000 years ago. Cary glacial activity in the southern extremity of the lobe is recorded during the period of 11,900 to 14,000 years B.P.

4. A part of the Des Moines lobe associated with the Altamont moraine in Iowa dated at 11,600 to 11,800 years B.P. antedates Two Creeks dated at 11,400 years B.P.

5. Post-glacial major vegetational and climatic changes in Iowa are recorded at 8,200 to 8,100 and 6,600 to 6,500 years ago. A gully cycle may be related to the latter major climatic change.

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Agronomy Building
Iowa State College
Ames, Iowa

U. S. Geological Survey
Washington 25, D. C.

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PALEOCLIMATE AS A MEASURE OF ISOSTASY

DANIEL I. AXELROD*

ABSTRACT. From Late Miocene well into the Early Pliocene approximately 2500-3000 feet of andesite was added to the 3000-foot summit region of the northern Sierra Nevada. If vulcanism actually increased the altitude of the range to nearly 6000 feet, then a major rain shadow must have developed to leeward.

Since sequences of Late Tertiary floras from western California, the west Sierran piedmont, and western Nevada show a similar decrease in rainfall occurred to leeward and windward of the range during and following the volcanic episode, vulcanism did not increase its height appreciably. Only on the east Sierran piedmont, where the floras record a greater reduction in rainfall, did a local rain shadow develop.

Paleoecological relations suggest that vulcanism increased the altitude of the range approximately 500 feet. This is consistent with the thesis that the Sierran block subsided isostatically under the accumulating volcanics, giving a net increase in altitude of 500-600 feet. When added to the 500- to 1000-foot ridge that already separated the lee side from the coast slope, it was sufficient to produce the local rain shadow, but not a regional one.

INTRODUCTION

The discovery of many Late Tertiary floras in the western United States in recent years has made it possible to direct more attention to the regional aspects of vegetation and climate, and to the influence that mountain ranges may have had on their distribution. In evaluating the effectiveness of topographic barriers, paleoclimatic and paleobotanic evidence has recently been used to estimate the height of the Sierra Nevada and western Nevada during Late Miocene time (Axelrod, 1956; 1957). Several lines of evidence have led to the conclusion that at the outbreak of vulcanism in the Late Tertiary, the Sierran summit probably stood at an average altitude of 3000 feet, the lowlands of Nevada to the east ranged from 2000 to 2500 feet, and that drainage was westward to the sea.

During the interval from Late Miocene well into the Early Pliocene, 2500-3000 feet of andesite accumulated along the summit of the northern Sierra Nevada, and thinned generally to the east and west. If vulcanism had actually increased the altitude of the Sierran crest to approximately 6000 feet by the beginning of the Middle Pliocene, a regional rain shadow must have developed to leeward. Evidence of it would be provided by a greater decrease in rainfall on the leeward than on the windward side. Reliable evidence of changes in precipitation on both sides of the range are provided by the continuous sequences of floras, or cliseres, from the Sierran province and bordering areas which span the volcanic episode. Rainfall estimates for the floras can accurately be made because they include species similar to living plants, and because the floras closely resemble modern forest, woodland, and scrub vegetation whose precipitation requirements are known. Rainfall trends for local areas can be established by analyzing sequences of successively younger floras. By comparing these cliseres on opposite sides of the Sierra Nevada, changes in rainfall and, by inference, in topography across the region can be reconstructed.

* This paper is based on recent studies in Tertiary paleobotany made possible by a grant from the National Science Foundation.

DISCUSSION

In terms of the present problem, four cliseres in the region from coastal California to western Nevada afford critical evidence with respect to the trend of Late Tertiary climate, and hence the relative effectiveness of the Sierran barrier just prior to, during, and immediately following the volcanic episode. Since these cliseres have been discussed in detail elsewhere (Axelrod, 1940; 1944; 1948; 1950; 1956), they need only be summarized here (tables 1-4).

TABLE 1

Cliseral Relations in Western California

Floras	Vegetation	Climate
Modern (lowland)	Oak woodland and savanna; chaparral; border-redwood forest	Subhumid to semiarid; rainfall from 17 inches in savanna to 25 inches at edge of forest; mild summers, cool winters
Mulholland & Petaluma (Middle Pliocene)	Madro-Tertiary woodland and chaparral dominant, with Arbutus, Arctostaphylos, Cercocarpus, Celtis, Fremontia, Lyonothamnus, Mahonia, Malosma, Photinia, Populus, Quercus, Rhamnus, Sapindus, Schmaltzia. Arcto-Tertiary forest on bordering slopes, with Acer, Alnus, Amelanchier, Castanopsis, Cornus, Fraxinus, Populus, Prunus, Rhus, Salix, Ulmus.	Semiarid, with subhumid aspect to forest; 18-23 inches rain in summer and winter. Summers warm, winters more mild than in area today
Black Hawk (Early Pliocene)	Arcto-Tertiary plants dominant, including Platanus, Populus, Ulmus. Madro-Tertiary plants rare, including Cercocarpus, Quercus, Schmaltzia.	Temperate, with 25 inches rain in summer and winter. Mild summers and winters.
Alamo (Basal Pliocene)	Arcto-Tertiary plants characteristic, with Betula, Persea, Platanus, Populus, Pterocarya, Smilax, Ulmus. Madro-Tertiary plants rare, including Quercus, Umbellularia.	Temperate, 30 inches rain, biseasonal occurrence with dry season less well marked than later in epoch. Mild winters and summers.
Loma Ranch (Late Miocene)	Arcto-Tertiary forest dominant, with Alnus, Berchemia, Castanea, Comptonia, Ilex, Persea, Platanus, Populus, Prunus, Nyssa, Taxodium, Smilax. Madro-Tertiary plants very rare, including Quercus, Umbellularia.	Temperate, with 30-35 inches rain well distributed through the seasons. Mild winters and summers.

The data provide clear evidence of the progressive decrease in annual precipitation in each area during the Late Tertiary. The nature of this change can be seen from the rainfall curves plotted in figure 1. The vertical bars on the graphs show, within a range of five inches, annual rainfall near the site of deposition as judged from the moisture requirements of related modern vegetation; the solid circle on each bar represents the average yearly precipitation at the site.

Plotting the Late Tertiary rainfall curves on a single graph (fig. 2) makes possible comparison of the changes in precipitation in each region. It

TABLE 2
Cliseral Relations on the West Sierran Piedmont

Floras	Vegetation	Climate
Modern (lowland)	Live-oak—digger pine woodland and savanna; grassland; scattered chaparral.	Semiarid, rainfall from 15 to 20 inches, increasing up slope; Hot dry summers, cool wet winters.
Oakdale (Middle Pliocene)	Madro-Tertiary plants dominant, with Arctostaphylos, Ceanothus, Celtis, Mahonia, Photonia, Quercus, Ribes, Robinia, Sapindus, Umbellularia. Arcto-Tertiary plants include only riparian species of Populus, Salix.	Semiarid, with 15 to 18 inches rain, in summer and winter. Summers and winters more mild than in area today.
Table Mountain (Late Miocene)	Arcto-Tertiary forest characteristic, with Acer, Berchemia, Carya, Cercis, Cornus, Crataegus, Nyssa, Persea, Platanus, Ulmus. Madro-Tertiary plants common, including Celtis, Cercocarpus, Forestiera, Ilex, Mahonia, Quercus, Robinia.	Temperate, with subhumid facies. Rainfall near 30 inches, as summer showers, winter rains. Mild summers and winters.
Valley Srings (Late Miocene)	Arcto-Tertiary plants prominent, including Acer, Alnus, Pinus, Platanus, Quercus, Salix, Torreya. Madro-Tertiary plants include Ceanothus, Lyonothamnus, Sapindus, Umbellularia in small numbers.	Temperate, with subhumid aspect. Rainfall from 30 to 35 inches, in summer and winter. Mild winters and summers.

indicates that there were normal climatic differences between these provinces, which ranged from sea level in western California, to a low piedmont within a few hundred feet of sea level in the Sierran foothill belt, to an altitude of approximately 2500 feet near the east margin of the range, and to 2000-2500 feet in western Nevada (Axelrod, 1957). The curves also demonstrate that there was a continuous, gradual decrease in rainfall in all areas into the Middle Pliocene, apparently the driest part of the Tertiary (Axelrod, 1948; 1950). In western California, on the west Sierran piedmont, and in western Nevada there was a decrease of 20 inches in annual rainfall, and at the same rate. From a paleoclimatic standpoint this means that only minor differences in topography continued to exist across the Sierran region during and following vulcanism. However, on the east Sierran piedmont in the immediate lee of the Sierran summit rainfall decreased fully 30 inches during the same interval, and at a more rapid rate. We may conclude that the piling up of volcanics was sufficiently high in this area to produce a low topographic-climatic barrier.

The significance of this local rain shadow can best be understood by recalling that at the inception of vulcanism the east Sierran piedmont was separated from the coastal slope by a low ridge closely corresponding in position with the present Sierran summit. Paleoecological evidence has suggested that at a maximum it probably stood not much more than 1000 feet above the Nevada lowlands which ranged from 2000 to 2500 feet. This inference

TABLE 3

Cliseral Relations on the East Sierran Piedmont

Floras	Vegetation	Climate
Modern (lowland)	Steppe (sage); piñon-juniper woodland; scattered yellow pine forest near at hand.	Semiarid, 10 to 13 inches precipitation as winter rain and snow chiefly. Cold winters, hot summers.
Red Rock (Middle Pliocene)	Riparian woodland of Populus and Salix, with grassland inferred for the interfluvial areas.	Semiarid, 12 to 17 inches rain, in summer and winter. Cool winters, warm to hot summers.
Verdi (Middle Pliocene)	Dominant riparian woodland of Populus and Salix, with scattered Madro-Tertiary oak-woodland (Quercus) and chaparral on nearby flats. Arcto-Tertiary forest with Abies, Arctostaphylos, Pinus, Populus, Prunus and Salix on bordering slopes.	Rainfall 18 to 20 inches in lowlands, increasing to 25 inches at edge of forest, chiefly in winter but some in summer; warm summers, mild winters.
Mansfield Ranch (Early Pliocene)	Arcto-Tertiary forest with Abies, Mahonia, and Quercus. Madro-Tertiary plants in small numbers, including Quercus.	Rainfall near 25 inches in lowlands, biseasonal. Warm summers, mild winters.
Chalk Hills (Early Pliocene)	Arcto-Tertiary forest dominant, with Abies, Amelanchier, Betula, Ceanothus, Pinus, Populus, Pseudotsuga, Quercus, Ribes, Salix, Sequoiadendron, Thuja. Madro-Tertiary plants rare, including Ceanothus, Garrya, Holodiscus, Quercus, Rhamnus.	Rainfall near 35 inches, distributed in winter and summer. Winters mild, summers warm.
Mohawk, Denton Creek (Late Miocene) interpolated	Arcto-Tertiary forest predominant, with Abies, Acer, Berchemia, Liquidambar, Persea, Pinus, Platanus, Quercus. Madro-Tertiary plants unknown.	Rainfall from 40 to 45 inches, in summer and winter. Mild warm summers and winters.

rests on the fact that in areas where the fossil floras show relationship to modern vegetation, ridges 1500-2000 feet high commonly result in marked differences in vegetation and climate from their windward to leeward sides. The broad similarity between the vegetation and climate of western Nevada and that on the west Sierran piedmont at the inception of vulcanism in the Late Miocene thus leads to the belief that the intervening barrier was not as high as 1500 feet. Paleoclimatic evidence suggests, therefore, that if we assume a maximum height of 1000 feet for the ridge separating western Nevada from the coastal slope in the Late Miocene, vulcanism could not have increased its altitude by much more than 500 feet; otherwise, a rain shadow of regional scope would have developed, the vegetation and climate would differ correspondingly, and both cliseres to leeward would record a greater reduction in rainfall than those to windward.

This conclusion based on paleoclimatic evidence agrees with geophysical data which suggest that the height of the range was not increased appreciably

TABLE 4
Cliseral Relations in Western Nevada

Floras	Vegetation	Climate
Modern (lowland)	Great Basin desert, with sage on middle slopes.	Desert to marginal steppe, 4 to 7 inches precipitation, as winter rain and snow. Summers hot, winters cold.
Truckee-Hazen (Middle Pliocene)	Small flora with Ceanothus, Juglans, Populus, Purshia, Quercus, Rhus, Salix of Madro-Tertiary origin chiefly.	Semiarid, with 15 inches rain in summer and winter. Warm to hot summers, mild winters.
Parran Junction (Early Pliocene)	Madro-Tertiary oak woodland dominant. Arcto-Tertiary plants rare, represented chiefly by Picea seeds.	Semiarid. 20 inches yearly rain in winter and summer. Warm summers, mild winters.
Fallon, Chloropagus (Late Miocene)	Madro-Tertiary woodland dominant in lowlands, with Arbutus, Cercocarpus, Juniperus, Quercus, Salix. Arcto-Tertiary forest on cooler slopes including Abies, Betula, Mahonia, Picea, Pinus, Salix, Sequoiadendron, Thuja.	Subhumid, with 25 inches yearly rain, increasing on bordering slopes, biseasonal distribution. Mild winters, warm summers.
Old Gregory (Late Miocene)	Arcto-Tertiary plants dominant, including Betula, Mahonia, Picea, Quercus. Madro-Tertiary plants in small numbers, chiefly live oak (Quercus).	Temperate. Approximately 30 inches rain, biseasonal occurrence. Mild winters and summers.

by vulcanism. For isostatic balance to be maintained, a load of andesite 2500-3000 feet thick would have to subside approximately four-fifths of this vertical distance, giving a surface only 500-600 feet higher than that prior to vulcanism. If this is added to the 1000-foot ridge which paleoclimatic evidence suggests may have been present, then the summit would have the inferred maximum height which might have existed without producing a regional shadow. At a minimum, the Sierran summit must have lain at least 500 feet above the Nevada lowlands prior to vulcanism, or a rain shadow of the magnitude recorded probably would not have developed during and following vulcanism.

CONCLUSION

Detailed paleoclimatic analyses of sequences of Late Tertiary floras in four areas extending from coast-central California across the Sierra Nevada and into Nevada, show that the climatic trend toward increased continentality was similar in all areas, that geographic position with respect to the ocean, mountains, and altitude produced consistent differences in local climates, and that comparisons of paleoclimates over the region provide a basis for measuring the amount of isostatic subsidence in the Sierra Nevada during vulcanism. Since the paleoclimatic results are internally consistent, and agree with that

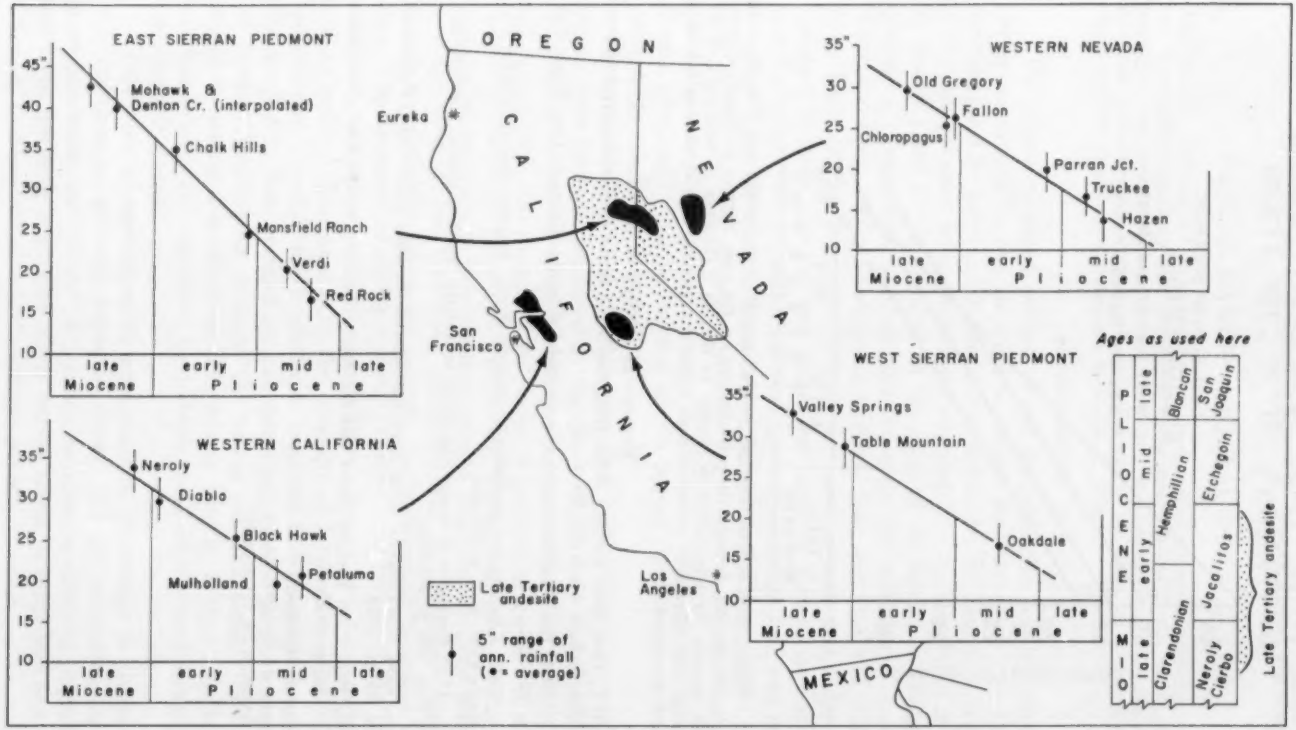

Fig. 1. Rainfall trends in California and Nevada inferred from paleoecological studies of fossil floras that lived during the following Late Tertiary vulcanism in the Sierra Nevada.

expected from geophysical data, we may conclude that our reconstructions of paleoclimate are based largely on sound principles of paleoecology.

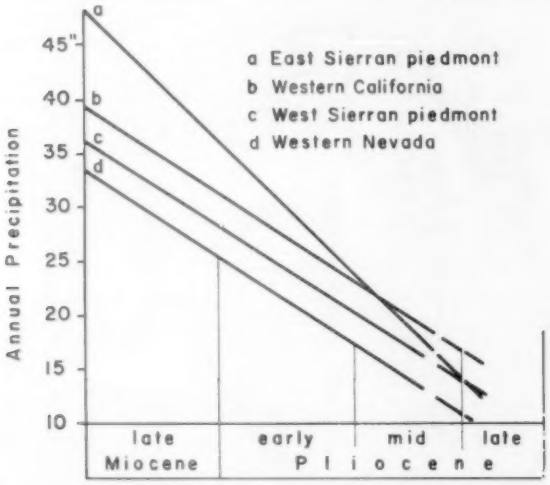

Fig. 2. Comparison of Late Tertiary rainfall trends in California and Nevada; data from fig. 1. The greater decrease in precipitation on the east Sierran piedmont as compared with western Nevada, which is similar to that in California to windward of the Sierra, means that vulcanism produced such a low topographic barrier that only a local rain shadow developed.

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Department of Geology
University of California
Los Angeles 24, California

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ARGON IN MICA AND THE AGE OF THE BERYL MT., N. H., PEGMATITE*

PAUL E. DAMON and J. LAURENCE KULP

ABSTRACT. A^{40} and K^{40} measurements on specimens of margarite (calcium mica) from two Appalachian localities and potassium micas associated with the Beryl Mt., N. H., pegmatite and associated country rocks are used to show the probable absence of appreciable primary argon in these minerals. The age of the Beryl Mt. pegmatite and associated regional metamorphism appears to be 323 ± 10 m.y.

INTRODUCTION

The potassium-argon method of age determination requires that the following conditions be met in order to obtain absolute ages: (1) The decay constant must be known; (2) the A^{40} produced by K^{40} decay during the life of the mineral must be quantitatively retained; (3) no significant A^{40} should be incorporated in the mineral at the time of its formation. Recent interlaboratory comparisons indicate that the analytical techniques are adequate (Carr and Kulp, 1957; Baadsgaard, Goldich, Nier and Hoffman, 1957). The decay constants are discussed below; they now appear reasonably well established by both physical and geological methods. Comparison of K^{40}/A^{40} ages on micas from a number of localities with uranium-lead and rubidium-strontium ages (Wetherill, Wasserburg, Aldrich and Tilton, 1956; Carr and Kulp, 1957; Goldich, Baadsgaard and Nier, 1957) has indicated that this mineral commonly retains all its argon, although there are cases where loss occurs, presumably due to thermal effects during the history of the mineral (Gast, Kulp and Long, in press).

It has been known for a half century that beryl invariably contains more helium than has been produced by the decay of uranium and thorium within the mineral subsequent to its formation (Strutt, 1908). More recently Aldrich and Nier (1948) have reported an excess of primary argon as well as helium in several essentially potassium free beryls. Wasserburg et al. (1956, p. 157) have called attention to this phenomena as a possible source of error in the K-A age method, but they have concluded that the available evidence, although not definitive, does not indicate the presence of inherited argon in mica. The discovery of excess (primary) argon and helium in the lattices of cordierite, tourmaline and some amphiboles (Damon and Kulp, in press), as well as in beryl, has pointed up the necessity of further investigation of this phenomenon in micas which otherwise appeared to be suitable for age work. The Beryl Mt., New Hampshire, pegmatite was chosen as the prime locality for further investigation because (1) excess gas had already been demonstrated in several minerals from this deposit (Strutt, 1911); (2) the deposit was young and, (3) age work by isotopic methods was very limited in this metamorphic district.

Two types of measurements were made in an attempt to define the possible existence of primary argon in micas: (1) A^{40} concentration in non-potassium micas and (2) A^{40}/K^{40} ratios on micas of differing potassium content from the same pegmatite where excess argon in beryl had been demon-

* Lamont Geological Observatory Contribution No. 267.

strated. The study of the non-potassium micas would show in general whether the mica-type lattice incorporates appreciable primary gas. The second study, although somewhat less sensitive, would show whether excess argon is present in potassium-micas in a normal pegmatite-country rock environment.

ANALYTICAL TECHNIQUES

The experimental techniques used in this laboratory to determine the potassium and argon content of minerals have been described by Carr and Kulp (1957) and Damon and Kulp (1957). Briefly, radiogenic argon (A^{40}) is released from the mineral in vacuum by direct fusion without flux. The radiogenic argon is equilibrated with a known amount of A^{38}, purified and completely recovered on charcoal at liquid nitrogen temperature, and the isotopic ratio is measured on a mass spectrometer to determine the amount of A^{40} present in the mineral.

Potassium analyses of the mica samples were made by Ledoux and Co. using a wet chemical method (Kallman, 1956). The potassium content of the albites was determined by flame photometry, and an upper limit was set on the potassium content of the beryls by a modification of the radiochemical method (Damon and Kulp, in press) first described by Gaudin and Pannell (1948). The analytical precision given for the results in table 1 has been determined from a large number of replicate measurements from previous work in this laboratory.

DECAY CONSTANTS

K^{40} decays by competing processes to A^{40} (K capture with associated gamma emission) and Ca^{40} (beta decay). At any time the fraction which decays to A^{40} is given by the ratio of the decay constant for K electron capture (λ_e) to the total decay constant ($\lambda_e + \lambda_\beta$) where λ_β is the decay constant for the beta transition. The amount of K^{40} which has decayed during a past time t measured from the Present is simply $K^{40}\ (e^{(\lambda_e+\lambda_\beta)t} - 1)$, and so the amount of A^{40} produced during time t is:

$$A^{40} = \frac{\lambda_e}{(\lambda_e + \lambda_\beta)} K^{40}\ (e^{(\lambda_e + \lambda_\beta)t} - 1)$$

or, solving for the time:

$$t = \frac{1}{\lambda_e + \lambda_\beta} \ln \left(1 + \frac{A^{40}}{K^{40}} \frac{(\lambda_e + \lambda_\beta)}{\lambda_e} \right)$$

Data on the necessary constants have been reviewed (Carr, Damon, Broecker and Kulp, 1956) covering the period up to late summer of 1955. The average of all reasonably accurate determinations from 1948 to the present, recalculated where necessary to conform to the latest knowledge of the decay scheme, gave $\lambda_\beta = 4.92 \times 10^{-10}$ and $\lambda_e = 0.574 \times 10^{-10}$ or 28.6 ± 0.9 dps/gm K and 3.33 ± dps/gm K for the beta and gamma activities respectively. Subsequently McNair, Glover and Wilson (1956) reported 27.6 ± .4β/g. sec. and 3.41 ± .10γ/g. sec. using proportional and scintillation spectrometry. Wetherill (1957), using refined scintillation technique and calibration against Co^{60} and Na^{24} gamma standards, obtained 3.39 ± 0.12

γ/g. sec. Backenstoss and Goebel (1955) reported 3.50 $\pm$ .01 γ/g. sec. using scintillation spectrometry and calibration with Pr^{142}, Co^{60} and Fe^{59}. For absolute geochronometry the uncertainties in these constants are still larger than is desired. By giving most weight to the most recent measurements of both beta and gamma determinations the authors have chosen 3.4 γ/g. sec. and 28 β/g. sec. for computation and consider them accurate to $\pm$ 5 percent. These correspond to $\lambda_e = 0.58 \times 10^{-10}$ and $\lambda_\beta = 4.8 \times 10^{-10}$ which will be used in this paper.

Wetherill et al. (1956) have calculated λ_e by a comparison of A^{40}/K^{40} ratios on minerals dated by concordant U-Pb and Th-Pb isotope ratios. They obtained $\lambda_e = 0.557 \pm .026 \times 10^{-10}$ assuming a value of $\lambda_\beta = 4.72 \times 10^{-10}$, the latter not being at all critical. This is a useful approach as the measured A^{40}/K^{40} age is more sensitive to λ_e in young minerals (<1. b.y.). Since the possibility of limited argon loss cannot be eliminated and the U-Pb ages for the three youngest localities used by Wetherill et al. are not certain to more than 3-5 percent, it seems better to use the most probable decay constants determined by direct physical means. Unless this is done, other potentially significant processes such as diffusion of argon from the minerals may be obscured.

The original data in this paper are given in terms of the A^{40}/K^{40} ratio because this is a directly measured quantity independent of the choice of decay constants.

ARGON CONTENT OF MARGARITE

Two specimens of margarite (calcium mica) were analyzed in an effort to detect lithospheric argon that may have been trapped in the mica structure at the time of mineral formation (table 1). These samples have a probable age of about 350 m.y.

TABLE 1

Argon Content of Two Specimens of Margarite

Locality	A^{40}* microliters per gram S.T.P.	K %	A^{40}/K^{40}
Ashland, N. C.*	<.001	<0.05	—
Unionville, Pa.**	~.004	~0.15	~.034

* 30 gram samples were fused. The A^{40} content has been corrected for atmospheric contamination (~.002 microliters per gram).

** from Emery and Corundum Co. Property, Ashland, N. C.

*** from Corundum Hill, Unionville, in the township of Newland, Chester Co., Pa.

Muscovite and biotite specimens of rocks of about this age contain .08 to .15 microliters per gram (Carr and Kulp, 1957). It appears from table 1 that margarite does not permit appreciable A^{40} to be built into the crystal at the time of mineral formation ($<1 \times 10^{-6}$ scc./g). A less likely interpretation is that the interlayer position in margarite permits loss by diffusion and thus an original excess of argon would not be detected.

ARGON CONTENT OF POTASSIUM MICAS AND OTHER MINERALS OF THE BERYL MT. PEGMATITE

The Beryl Mt. pegmatite has been intruded into the schistose amphibolites and interbedded metasediments comprising the Littleton formation about 0.8 miles south of the village of South Acworth in the Keene pegmatite district of New Hampshire (Cameron et al., 1954). A sketch map of the Beryl Mt. area is shown in figure 1. Most of the minerals studied came from the main pit

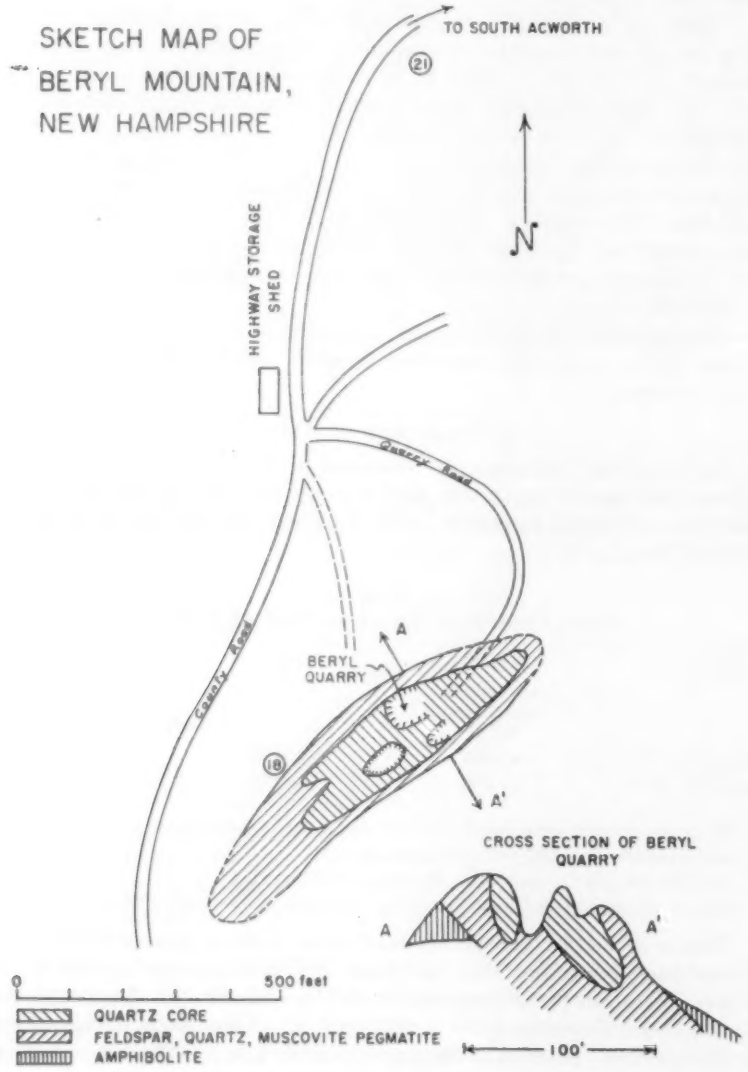

Fig. 1

where the pegmatite is best exposed and in which beryl is abundant. The pegmatite has a core of massive white and rose quartz. The core is flanked by an intermediate zone consisting primarily of perthite and quartz which contains large cylindrical crystals of albite, large books of wedge mica and crystals of green and yellow-brown beryl up to 5 feet in length. The intermediate zone, which is about 25 feet to 50 feet wide, grades into a wall zone consisting primarily of coarse graphic granite but containing small clusters of biotite, sooty black crystals of tourmaline and small garnet dodecahedrons. The origin of the pegmatites in the Keene district, according to Cameron et al. (1954), is either from underlying bodies of the New Hampshire magma series (Concord granite), or from the concordant injections of Bethlehem gneiss that occur abundantly in the Littleton formation.

TABLE 2

Argon in the Beryl Mt. Pegmatite

Sample Number	Mineral	A^{40*} microliters per gram S.T.P.	K %	A^{40}/K^{40}	Apparent Age m.y.
A. *Micas*					
BMD-16m	muscovite	0.120	8.39	0.0209	329±13
		0.123	8.70	0.0208	328±13
BMD-3b	muscovite	0.118	8.48	0.0203	321±13
BMD-15	muscovite	0.0995	7.22	0.0202	319±13
BMD-10	muscovite and biotite from xenolith	0.0969	6.88	0.0206	327±13
		0.0879	6.87	0.0188	299±13
BMD-18a	biotite from contact zone	0.0696	6.47	0.0160	258±10
		0.0735	6.69**		
BMD-21	impure biotite from Littleton	0.0245	1.67	0.0215	338±15
			Average (18a omitted)		323±6
B. *Other Minerals*					
BMD-16p	perthite	0.0897	10.4	0.0136	238±10
BMD-3a	albite	0.0074	0.39	0.026	402±94
			0.42**		
BMD-3d	albite	0.0064	0.37	0.025	390±85
BMD-9	beryl	0.045	<0.05	>1.3	
BMD-2d	beryl	0.056	<0.1	>0.8	

* Radiogenic A^{40} after correction for atmospheric contamination. The correction varied from a few percent for muscovite up to 20 percent for albite.

** Analyzed by E. Oslund and C. O. Ingamells of the Univ. of Minnesota, Rock Analysis Laboratory.

Decay Constants $\lambda_e = 0.58 \times 10^{-10}$ yr.$^{-1}$; $\lambda_\beta = 4.8 \times 10^{-10}$

Analytical precision:
micas and perthite: $A^{40} = \pm 3.3\%$ s.d., $K = \pm 3\%$ s.d.
$A^{40}/K^{40} = \pm 4.3\%$ s.d.
albite: $A^{40} = \pm 20\%$ s.d. (est.); K determined by flame photometry (S. Kallman—Ledoux, mc) $= \pm 10\%$; $A^{40}/K^{40} = \pm 22\%$.

The results are given in table 2. The measured ages of three book muscovite samples from the main pit (BMD-16m, BMD-3b and BMD-15) and a mixture of muscovite and biotite separated from a xenolith of country rock enclosed in the pegmatite (BMD-10) are in excellent agreement despite a difference in potassium content of 25 percent. The agreement of the pegmatite and xenolith ages also shows that the country rock mica does not have any "inherited" argon from its existence prior to pegmatite intrusion. The beryl sample BMD-9, which like BMD-2d shows a large quantity of primary argon, was in direct contact with muscovite BMD-3b. Even in this case the muscovite shows no evidence of excess argon.

Biotite (BMD-18a) taken from the contact zone between the amphibolite country rock and the western side of the pegmatite shows an anomalously low age. X-ray examination of this specimen shows it to be partly chloritized which probably accounts for the low argon content. Further evidence of alteration was the unusual quantity of water and other volatiles released during fusion of this sample.

The perthite sample (BMD-16p) shows the usual (Wetherill, Aldrich and Davis, 1955; Carr and Kulp, 1957, Goldich, Baadsgaard and Nier, 1957) loss of argon. The albite samples appear to give the correct age, although the errors are large due to the low potassium content. The apparently higher retention of argon in albite compared with perthite is in contrast with the observation of Wasserburg and Hayden (1956) but may be merely the result of included veinlets of sericite (fine muscovite) which may contribute most of the potassium content.

AGE OF THE BERYL MT. PEGMATITE AND SURROUNDING ROCKS

The potassium-argon age of biotite separated from the Littleton formation about a quarter of a mile from the Beryl Mt. pegmatite was 338 $\pm$ m.y. which agrees with the pegmatite age within the experimental error. Table 3 compares the K/A ages from the Beryl Mt. pegmatite with lead-alpha ages on zircons, xenotime and monazite for the major units of metamorphic rocks in the area.

TABLE 3

Comparison of A^{40}/K^{40} Age of Beryl Mt. Muscovites with Pb/α Ages for the Oliverian and New Hampshire Magma Series

Geologic Formation	Age m.y.	Method and number of Analyses	Analyst This paper
Beryl Mt. Pegmatite	323	A^{40}/K^{40}: ave. for 5 micas	Lyons*
Concord Granite	330	Pb/α: ave. for 2 zircons, 1 xenotime and 1 monazite	Lyons*
Bethlehem Gneiss	317	Pb/α: ave. for 1 monazite and 1 zircon	
Oliverian magma series	322	Pb/α: ave. for 3 zircons	Lyons*

* See Faul (1954) p. 268.

It appears that the last major metamorphic event in this region took place about 325 m.y. ago, that the pegmatites and the Concord granite were

formed at the time of the regional metamorphism rather than later and that they are probably earlier than the Upper Devonian-Carboniferous age suggested on geologic grounds by Cameron et al., 1954. In any case the potassium-argon date on the Beryl Mt.-Littleton rocks places an upper limit to the time of deposition of the Littleton formation. If the Holmes-Marble (Marble, 1950) time scale is accepted, these results would make the Littleton deposition and metamorphism pre-Devonian. On the other hand if the Littleton deposition is taken as Oriskany (Lower Devonian) in age as the fossil assemblage indicates (Billings, 1937)—a more plausible approach—this may be a new and important control point in establishing the absolute post-Cambrian time scale. It would then be concluded that the Silurian-Devonian boundary is at least 330-340 m.y. ago.

Appendix

Description of Beryl Mountain, N. H., Pegmatite Samples

BMD-2d	Beryl, surrounded by 1/4" mica, smoky quartz and a large albite crystal. See Damon and Kulp (in press).
BMD-3a	Albite associated with quartz and beryl containing sericite veinlets near contact of quartz core with quartz-mica-feldspar intermediate zone.
BMD-3b	Large books of muscovite about 6 inches wide associated with quartz, albite and beryl; sample was carefully hand-picked to remove all visible impurities.
BMD-3d	Albite similar to BMD-3a.
BMD-9	Beryl, in contact with BMD-3b. See Damon and Kulp (in press).
BMD-10	Mixed muscovite and biotite from xenolith of quartz-mica schist containing nearly equal amounts of clean, unaltered, biotite and muscovite; garnet occurs as an accessory; one patch of tourmaline parallel to schistosity was visible in thin section; sample was pure except for a few small crystals of garnet.
BMD-15	Muscovite from graphic granite; several mms. in width; sample was separated by sieving and purified by hand-picking; obtained from main beryl pit in area where the intermediate pegmatite zone grades into the coarse graphic granite.
BMD-16m	Muscovite samples from a large book about 12 inches wide and 6 inches thick was free of inclusions and associated with perthite and quartz.
BMD-16p	White to honey-yellow perthite adjacent to BMD-16m containing a small amount of sericite in thin veins visible in thin section.
BMD-18a	Biotite from contact zone between western pegmatite and amphibolites near county road (fig. 1). Hand specimens nearest the contact contained about 40 percent hornblende, 15 percent biotite; remainder was primarily quartz and feldspar. Thin section of rock nearest contact, where biotite sample was obtained, showed that the grains were not well aligned but, rather, occurred in swirls and eddies with evidence of flow. Thin section from rock 10 feet from the contact showed well aligned, finer grains, no evidence of movement, no biotite but considerable fine-grained clean muscovite, much more quartz, less hornblende and scattering of opaques. Because of the coarseness of the biotite at the contact there was no difficulty in removing the hornblende by sieving. Remaining impurities were removed by heavy liquid separation (tetrabromethane) and water washing (panning).
BMD-21	Impure biotite concentrate from country rock, a quarter of a mile N from the pegmatite in a pasture beside the county road (fig. 1). It was similar to the amphibolites near contact of the pegmatite, but contained less amphibole; quartz and feldspar were the principal constituents; small amount of garnet. Sample was prepared by sieving; sericite was present but was removed by the sieving; heavy liquid separation (tetrabromethane) and water washing were used to remove amphibole. Later examination showed that considerable impurities remained, including amphibole, which contribute, in part, to the low potassium content.

CONCLUSIONS

1. The absence of measurable primary argon in calcium micas or in sets of potassium micas of differing potassium content suggests that primary argon is not a significant source of error in the potassium-argon dating method.

2. The time of regional metamorphism and pegmatite intrusion in the region near Beryl Mt., New Hampshire appears to be 323 $\pm$ 10 m.y. ago.

ACKNOWLEDGMENTS

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LAMONT GEOLOGICAL OBSERVATORY
COLUMBIA UNIVERSITY
PALISADES, NEW YORK

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THE DIRECT RECORDING OF SOLIFLUCTION MOVEMENTS

P. J. WILLIAMS*

ABSTRACT. A new instrument has been developed for the direct recording of solifluction and similar small mass movements. It also gives some indication of the vertical velocity profile (relative movement at each depth).

Probes, which are inserted into the soil, consist basically of spring steel strips with electrical resistance strain gauges attached. Bending of the probe by soil movement can easily be determined at any time from the strain gauge readings.

The probes can be constructed sensitive to very small soil movements. They do not have any appreciable effect on the mechanics of the soil. The apparatus is simple and fairly cheap to construct, but its sensitivity and accuracy are to some extent dependent on the care with which the instrument is constructed and used in the field. In a field investigation of solifluction below a late snow patch, interesting information has been obtained concerning the depth, nature and timing of soil movements, which may be compared with environmental factors suspected of causing the movements.

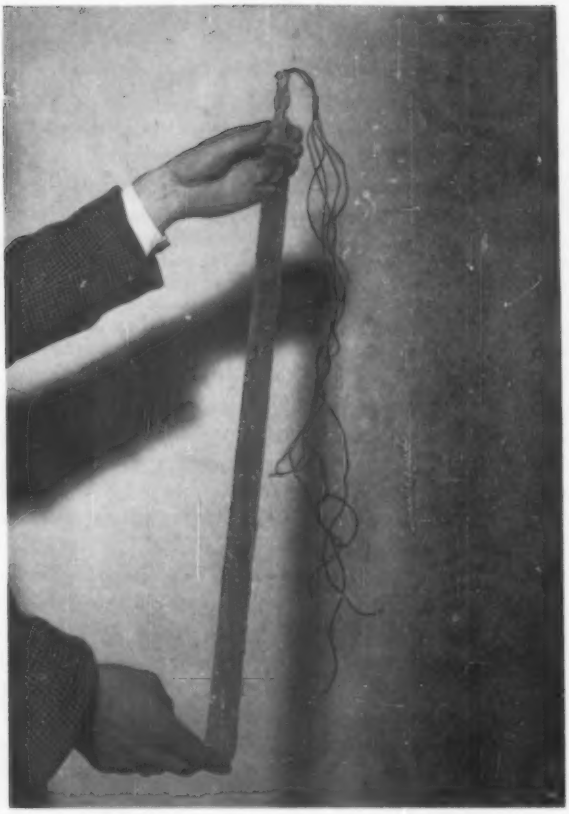
In experimental studies of solifluction there has long been a need for an instrument which will record soil movements as they occur, so that they may be compared with other changing factors possibly causing them. An instrument has been developed during the past two years for this purpose, and it has proved to give valuable information in investigations of the mechanics of a solifluction slope. It also gives some indication of the vertical velocity profile (relative movements at different depths). The instrument could be used in other geotechnical or glaciological investigations where fairly small soil or ice movements may be profitably investigated.

In principle, the instrument consists of probes (plate 1), sensitive to bending, which are inserted into the soil under investigation. The determination of the curvature set up in an inaccessible rod or pipe of small diameter is an old problem, and several methods were considered. I am indebted to Prof. Dr. Ivan Rosenqvist for his suggestion of the use of electrical resistance strain gauges. The probes are constructed basically of spring steel strips of convenient length, for example 1 to 2 m long, about 2 cm wide and about 1 mm thick. The strain gauges are attached to these strips and the strains set up by bending of the probe are recorded.

The gauges are of the now well-known type, which consist of an element, or grid, of wire, about 2 cm long and 3 mm wide, attached to a paper backing, whose electrical resistance, determined by a sensitive measuring bridge, is approximately proportional to the strain set up in the element. The strain set up in the surface of a strip by bending, as shown in figure 1a, is proportional to the deflection d, provided that there is no twisting of the strip. This strain is recorded by the gauges which are attached to the strip parallel to its length (fig. 1b). To gain a picture of the form taken up by the probe during soil movement, and hence of the vertical velocity profile (which will not be simply the arc of a circle), the strain must be determined at a number of points. There are several good accounts of the theory and general usage of resistance strain gauges (e.g., Yarnell, 1951), but this use of them in wet soil under field

* Present address: Division of Building Research, National Research Council, Ottawa, Canada.

PLATE 1



Short probe. (photo, Magne Arneberg).

conditions poses a number of special problems. Furthermore there are certain limitations to the results that can be obtained:

1. The strain gauge readings only give a quantitative measure of the curvature of the strip over the length where they are attached. Strain determinations over the entire length of the strip would be necessary to determine precisely the form taken up by the probe, and this would involve the use of an impossibly large number of gauges and leads, especially in an investigation with a number of probes. It is found that 2 to 3 gauge pairs per meter length of probe are most satisfactory under average conditions, but there is then an element of conjecture in any calculation of the vertical velocity profile. The possibility of considerable strain changes due to pressure from for example, an adjacent stone, being mistaken for general bending of the probe and a general soil movement, is prevented by the use of two or more probes some distance apart.

2. Where the movement is restricted to a shear plane, or thin layer of soil, the form taken up by the probe cannot represent the vertical velocity profile.

3. The forces due to soil movement acting on the strip are unlikely to be limited to the one plane parallel and normal to the gauges on the strip, so that to some extent the strain readings will be a result of twisting of the strip.

4. There may be a gradual change of the calibration values of the gauges under operation, because of hysteresis, or of drift due to various causes.

5. A high degree of accuracy cannot be obtained in the measurement of gauge resistances under field conditions.

6. Very small movements of the soil may be resisted by the probes.

These points are discussed further below.

CONSTRUCTION OF THE PROBE

Temperature variations produce a change of resistance which is compensated for by the use of pairs of gauges, the second gauge being placed at the same level on the opposite side of the strip. In the measuring bridge circuit the gauges are in opposite arms, and, provided they undergo similar temperature changes, the temperature effect is then balanced out.

The gauges are affixed to the steel strip by the use of special adhesives. There are normally two leads from each gauge, but the return from all gauges on a strip may be made through the strip itself. In this way, a probe with 3 gauge pairs will have a total of 7 leads. In an investigation involving the use of several probes a further economy may be affected by combining the return leads of several probes, especially if the measuring bridge is situated some distance from the probes.

The gauges must be protected from physical damage by, and electrically insulated from, their surroundings. There must not be any humidity variations around the gauges. These problems seem to have been overcome after trials with a variety of materials by enclosing the entire probe in a sheath of plastic. A cold-setting plastic of the Araldite type is used. This is obtained as a translucent, syrup-like material, which on the addition of an acid hardener, sets after a short time to give a flexible plastic of considerable strength, with a fairly smooth surface, and a high degree of chemical inertness. It is also an excellent electrical insulator. After the gauges and their leads are attached to the steel strip, a single layer of gauze bandage is carefully wound round the whole. The fluid plastic, with the hardener added, is then liberally coated over the bandage. After the first application has set, further coats of the plastic may easily be applied until the plastic sheath is judged to be complete and strong enough (when for example the plastic is 3 mm thick). A further precaution is to apply a small quantity of the plastic directly on to the gauges and lead connections. This provides additional protection to the gauge, but unless this coating is very small and not directly attached to the outer sheath, further strains may be set up in the gauge, which are not directly proportional to the curvature of the strip, and in extreme cases the gauge may be wrenched from it. The gauge readings then correspond roughly to the curvature of the probe, but accuracy is greatly reduced.

SENSITIVITY OF THE PROBES

The strain set up by the curvature of a thick strip or beam (fig. 1a) is given by

$$e = \frac{4dh}{L^2}$$

where h = thickness of the strip, and d = deflection at the center between two points distance L apart.

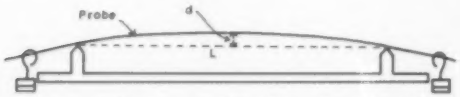


Fig. 1. a. Apparatus for calibration of probe.

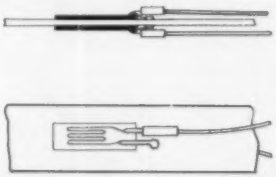


Fig. 1. b. Attachment of gauge to strip (approx. natural size). Side view above, plan view below.

It has been found in a series of laboratory trials that the observed curvature of the thin strips, and of the probes, for a given strain change is slightly more (for example, by 10%) than that calculated, but the relationship between curvature and strain is linear, and the probes may easily be calibrated empirically. The probe is arranged as shown (fig. 1a) and the deflection d, under loading, measured optically. The addition of the plastic sheath also results in a certain elastic lag, such that if the unloading of the probe is completed within a few minutes the original state of strain is not immediately regained. It is unlikely that this has any significance in the use of the probes in the field.

The very small curvature of a strip a meter long of thickness 1 mm, where the deflection at the center of the strip is 1 mm, gives a strain of about 4 microstrain.[1] This corresponds to a resistance change approaching the smallest measurable in a field layout. It is clear that a very small curvature of the probes can be detected.

The relationship given above indicates the importance of the thickness of the steel strip in determining the sensitivity of the probe. The strain produced by curvature of the strip is approximately proportional to its thickness, but if the strip is too thick the probe will have a considerable resistance to bending. The question arises of whether the probes have any significant strengthening effect on the soil, or whether the soil will tend to move round them rather than bend them. In either case the probes would not give a true indication of the soil movements. Consideration of the mechanics of the soil indicate that only in certain cases is this liable to become important. Movement occurs by shearing of the soil. Shearing will only occur when the resistance to shear (or shearing strength), of the soil has a smaller value than the shearing stress, i.e. the sum total of forces tending to produce shear (see any textbook on soil mechanics; e.g., Taylor, 1948). When movement occurs in a previously stable slope, this may be either because of an increase in shearing stress (e.g., by an

[1] One microstrain corresponds to a strain of $e = 10^{-6}$.

increased load on the slope) or because of a decrease in the shearing strength (e.g., as a result of saturation of the soil by water), or by both effects simultaneously. It has been shown that the changes in shearing stress and strength that can occur seasonally in a solifluction slope are large (Williams, 1957). In cases where the shearing strength of the soil is fairly large, the slight added component of strength, due for example to a strip 1 mm thick and its plastic sheath is not likely to hinder movement appreciably or to delay its onset. Only in extreme cases where the shearing strength is extremely low (and where therefore the soil must be on the very slightest slope) can the probe possibly represent an important component in the strength of the soil, and therefore in the value of the critical shearing stress necessary to cause movement.[2] A similar argument may be used to show that in general the moving soil will carry the probe with it rather than flow round it (which would involve further shear), i.e. that the curvature taken up by the probe is a fair representation of the relative movements at different depths (vertical velocity profile). These considerations may be summarised by saying that in general the forces operative when movement occurs have a completely different order of size from those necessary to bend the probe.

It is obvious that wherever any definite shear plane is developed, or whenever the soil movement is restricted to a relatively thin layer, the probe cannot take up a form representing the vertical velocity profile. In such cases the onset of shear will be indicated by the probe (which also may well show the gradually increasing flow-like movements which often precede shear along a definite plane). It could be expected that some change of strain gauge reading would occur for a period after the main soil movement has ceased, while the probe takes up an irregular curve adjusted to the new position of the soil. In many cases of solifluction however, soil movement is not restricted to a shear plane or thin layer. Provided that this is verified as far as possible by careful observation, and by consideration of the type of solifluction, then the bending of the probe can be assumed to occur with the movement and in relation to the vertical velocity profile.

In practice, strips of about 1 mm thickness and 2 cm width are found to give probes of sufficient rigidity to be conveniently handled out of the ground (weaker probes may sag under their own weight so that the gauges may be overstrained), and of sufficient sensitivity, while in general not possessing sufficient rigidity for them to resist or hinder movement by the soil.

A portable battery-operated bridge is necessary for determining the resistance of the gauges in the field. There are several bridges of different sensitivity especially made for strain gauge measurements. The sensitivity required will depend on the fineness of the movements to be investigated, and the stability of the other apparatus. In field applications the accuracy of the gauge readings is to a considerable extent dependent on the number of connections and their quality. All external connections should be soldered with acid-free flux and completely enclosed in a plastic sheath. Moisture inside the insulation of poorly completed connections is sufficient to set up a small

[2] Such low shearing strengths may occur in the special case of saturated soil on approximately level ground, where the internal stresses are entirely borne by water pressure.

E.M.F. such that a deflection of the galvanometer needle can be obtained, for example, by squeezing a connection between the fingers.

Readings of the strain gauges are taken at convenient intervals, every hour if the soil is in an actively moving state, or perhaps only daily if the soil is apparently stable. A series of hourly readings is shown in figure 2. The lower line represents a probably stable soil, and the upper line shows solifluction movements. It will be seen that there is a considerable scatter in the results. This is probably not due to changes of strain (as might be caused by very slight soil disturbances), but to imperfections in the circuit of the type mentioned above.

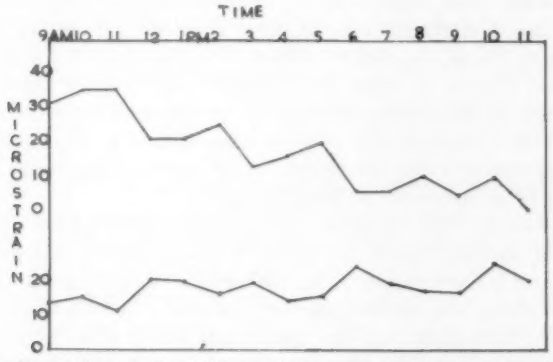


Fig. 2. Readings taken during 14 hours from probe with two gauge pairs. The upper gauge pair (upper curve) clearly shows solifluction movements; the lower gauge pair is in frozen soil and shows little change.

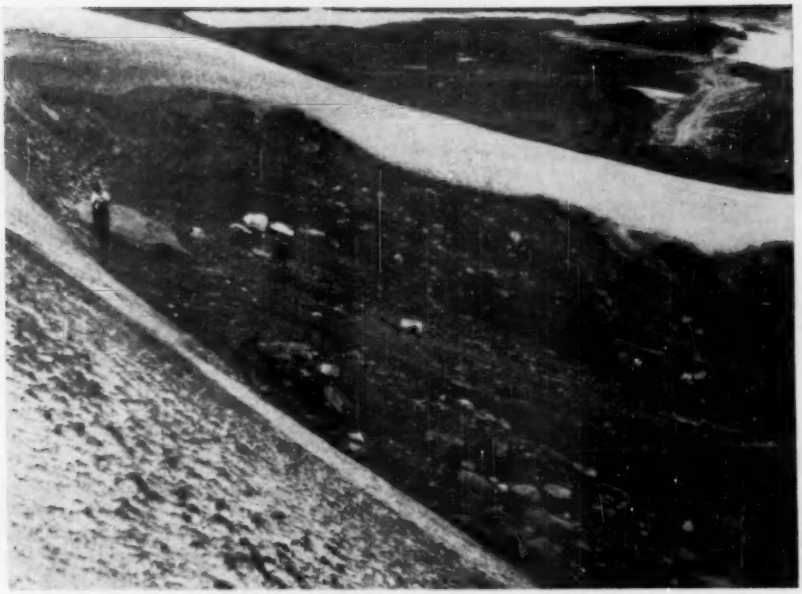
FIELD USE OF THE PROBES AND INTERPRETATION OF READINGS

After field trials of these probes, a detailed investigation was carried out during 4 weeks in Dovrefjell, Norway. The slope chosen for this investigation is shown in plate 2. Solifluction was suspected on the slope beneath the late snow patch (the picture was taken in early July).

A field layout was set up over about 2,000 sq m, with 10 strain gauge probes 1 to 2 m long, 17 thermometers measuring temperatures at various depths, and 5 piezometric tubes measuring pore-water pressures. Air temperatures were recorded with a thermograph. A series of general observations of the slope, snow patch, soil density, etc., were also made. In this way the occurrence, timing and size of soil movements could be compared with a variety of factors likely to cause or influence them. A discussion of the results of the investigation and of the mechanics of the slope is given in Williams, et al. (1957); summarized in Lewis, 1957), and only a short account of the results from readings of the probes will be given here.

In order to avoid undue disturbance of the slope, and to record possible soil movements throughout the thaw, the probes were inserted into the soil while it was still frozen. Holes were made in the frozen ground with a portable pneumatic drill. This made holes of slightly larger diameter than the breadth of the probes, and it was found that as long as the air temperature

PLATE 2



Site of the investigation of solifluction described. (photo, D. R. M. Lillistone).

was above freezing the drill operated quite efficiently. Some difficulty was experienced when the air temperature was several degrees below freezing, because the bit tended to become frozen in should it cease to rotate for a short while. After insertion of the probe into the hole, mud carefully freed of stones was poured in to fill the larger spaces.

Leads from the probes were run to a tent placed at one corner of the layout (see plate 2), where the measuring bridges were housed. To achieve consistent results it is necessary to have the leads into the bridge permanently attached, and to select the gauge pair to be read by switches. A switch unit containing high quality wafer switches of high resistance constancy was constructed for the purpose.

Disturbance of the soil while the observations were in progress was avoided by reading the mercury thermometers from afar, with a theodolite. Whenever any instrument in the layout required adjustment, it was approached by walking on still-frozen ground, the snow, or large boulders.

Two of the probes recorded very considerable solifluction movements, three showed probable small movements, and two probes on presumably stable parts had approximately constant readings. The results for the remaining probes were complicated by modifications under trial.

Figure 3 shows graphs of the readings taken at intervals during three weeks from a 1.5 m long probe with two gauge pairs, together with an inter-

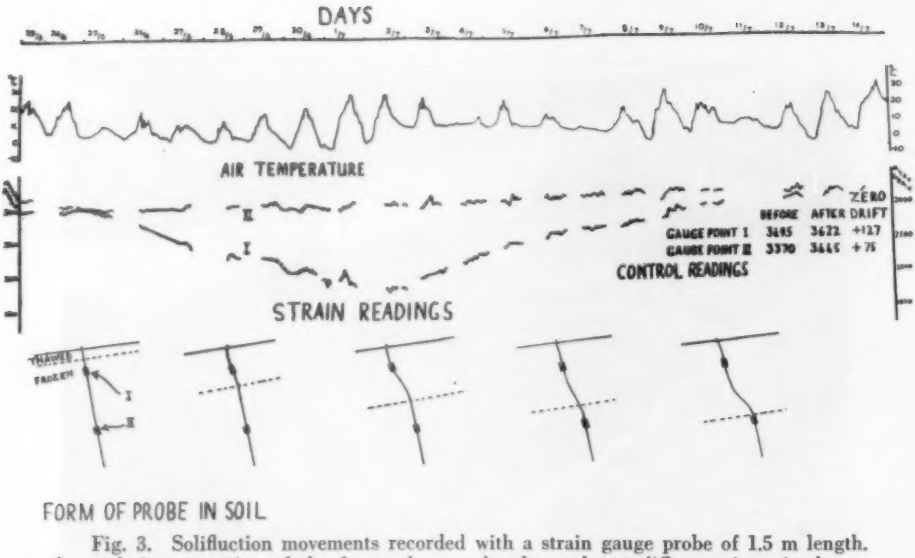


Fig. 3. Solifluction movements recorded with a strain gauge probe of 1.5 m length. A rough interpretation of the form taken up by the probe at different times is shown. The latter is based on the assumption, supported by the observations of several probes in the area, that the soil movements occur rather regularly, and progressively at increasing depth.

pretation of the forms taken up by the probe, corresponding to the readings at given times.

In the determination of the vertical velocity profile, it is assumed that each strain gauge pair lies on the centre point of an arc of a circle. The curvature of the arc at a given moment is determined from the strain reading according to the relation given on page 708. The profile is then drawn up as two or more adjacent arcs, according to the number of strain gauge pairs on the probe. The gauges give the exact curvature only for their own length, and the profile is not generally a series of distinct arcs, so that it is finally necessary to draw in the most likely form for the profile on the basis of the arcs and earlier and later readings.

Calibration control readings were taken before and after the investigation to determine the size of the drift. Readings of the gauges were taken with the probe aligned in a marked position against a steel pipe. Although this procedure is rather crude, in the majority of cases a near correspondence was obtained. The average drift over the 4-week period, in the 15 gauge pairs of the 7 probes named above, was a resistance change of the same size as would correspond to 78 microstrain. 7 gauge pairs however, had a drift corresponding to 35 microstrain or less, and only one pair had a drift corresponding to over 150 microstrain. The strain set up by the curving of a strip 1 m long and 1 mm thick, with a centre deflection of about 20 mm is 78 microstrain. A deflection of about 9 mm corresponds therefore to 35 microstrain. In the inter-

pretation of the readings the drift effect must be taken into account and if necessary a correction applied. The drift effect is roughly dependent on the interval of time between control readings, and if considerable accuracy is essential, the probes must be controlled more frequently. In other cases it is found, as in the example shown, that the drift is small in relation to the resistance (strain) changes due to curvature of the probe.

As noted above, for the correct interpretation of results obtained from the use of the probes, careful consideration must also be given to the nature of the movement under investigation. Very thorough observations were made of the soil in this area of solifluction. There was no indication of any definite shear planes in the area where the probes were placed, nor of any thin layers where movement might occur preferentially. Before the probes were removed, their alignment in the soil was reobserved as far as possible, as a further check, and where considerable movement had occurred (as in the example, fig. 3), the change was conspicuous.

The completed study enables the following conclusions to be drawn:

1. In the spring of 1956, which can be regarded as a typical one with respect to weather, solifluction movements occurred only locally over this slope.

2. The movements may persist to a depth of more than 75 cm, while points on the surface may move downslope more than 20 cm.

3. Soil movements generally do not occur suddenly, but continuously over a period of days or weeks. There is no marked diurnal or other rhythmical variation.

4. At any one time, it is likely that the soil in the region nearest the still frozen ground moves fastest, although the vertical velocity profiles observed over a period of time show some variation from this.

5. Provided its limitations are understood, this instrument can give useful information in the study of soil movements and their timing.

ACKNOWEDGMENTS

I am grateful to Professor Rosenqvist for his suggestion of the use of strain gauges, and for a visit to Norges Tekniske Höyskole, where I was initiated in their use. The instrument was constructed while I held a Norwegian State Scholarship.

The field work was a part of the program of two expeditions of willing Cambridge and Oslo students, of whom Mr. Eric Smith deserves especial thanks for his skill and care in the construction of much electrical apparatus. The expeditions were supported by the Royal Geographical Society, the Norwegian Nansen Fund, the North Sea Foundation, the Norwegian Army and others, including Messrs. H. Tinsley & Co. who loaned a portable strain gauge bridge. To all of these thanks are due.

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Geografisk Institut
University of Oslo
Blindern, Oslo, Norway

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NATURAL RADIATION DAMAGE IN ALBITE

JOHN McANDREW

ABSTRACT. Uraninite inclusions in low-temperature albite from the Cloncurry District, North-West Queensland, are surrounded by prominent extinction haloes in the albite. The haloes arise from radiation damage of albite by the α-particle bombardment, which produces a change of 42° in 2V, and a 48° rotation of the optical indicatrix about an axis close to the crystallographic direction [4.3 $\bar{1}$ 2.2]. The difference between this change and the change to high-temperature albite is in accordance with the different mechanism of structural disorder involved.

INTRODUCTION

The common effect of radiations from radioactive decay on minerals sensitive to these radiations is the production of color haloes which may be pleochroic, or a disruption of the crystal structure with decreasing density and refractive index.

The major visible effect of these radiations on albite is, however, a change in the optic axial angle with simultaneous rotation of the optical indicatrix, thus producing a change in the extinction position. In thin section such alteration produces a halo which is visible between crossed polars by extinguishing at different positions to the unaltered albite, the change in extinction increasing progressively towards the boundary of the uraninite.

In ordinary light no change is visible in the albite other than some fractures sub-radial to the uraninite. The haloes are referred to therefore as extinction haloes.

OCCURRENCE

Extinction haloes around uraninite inclusions in albite are strikingly developed in thin sections of albite rock from a mineral lease called Clear Waters, which is approximately 1½ miles S S E of the 34 mile peg from Cloncurry on the Cloncurry-Mount Isa road, Queensland, Australia (McAndrew, 1957).

The radioactive albite rock is fine grained and mottled in cream to reddish brown colors in hand specimen. An ill-defined banding is visible only in thin section, by variation in grain size of the albite, and in the occurrence of minor constituents. Albite-hornblende rocks occur in the same beds to the north and south in association with scapolite (Edwards and Baker, 1954).

Albite is the major constituent, with smaller amounts of hornblende, calcite, quartz, uraninite, sphene, magnetite, hematite, apatite, chlorite, a mineral comparable with uranophane, and pale green and brown alteration products of uraninite.

Thin sections reveal that the *albite* consists of fine grained aggregates in which crystals are about 0.04 mm across, with irregular elongated patches of coarser grained albite in which individual crystals are up to 0.6 mm across. Many of the smaller, and a few of the larger crystals are untwinned, some are crushed, or have curved twin planes. The refractive index, $n_\gamma = 1.530 \pm 0.002$, and the 2V of −77° and the optical orientation, determine the albite as the low-temperature variety.

Hornblende occurs in variable amount, as subhedral crystals which have a grain size similar to the adjacent albite. *Calcite* occurs irregularly as crystals up to 1.5 mm across, some enclosing albite in a poikiloblastic texture. *Quartz* is intergrown with albite as anhedral crystals of about the same size as the albite, and rarely up to 2.5 mm across.

Occasional crystals up to 0.05 mm across are identified from optical properties as *uranophane*. Some narrow veinlets, and films along grain boundaries and cleavages of the albite, consist of a secondary mineral which has similar optical properties, as far as can be ascertained, and is tentatively identified as uranophane.

Uraninite is the most abundant opaque mineral, and is seen in polished section as grey, isotropic, subhedral crystals which are up to 0.13 mm in diameter, though mostly smaller. The cell-edge of uraninite which shows no visible alteration was determined from an *x*-ray powder pattern as 5.47 Å. As no thorium has been detected in the specimens (Dr. T. R. Scott, personal communication), this cell-edge indicates that the fresh uraninite is unoxidized, and of composition close to $UO_{2.0}$ (Brooker and Nuffield, 1952). The uraninite is concentrated in bands in the rock. Some shows alteration by weathering, firstly to a substance which is darker grey in reflected light and has yellow-brown internal reflections. Further alteration produces the green and brown transparent, isotropic residuals observed in thin section.

Other opaque minerals are euhedral magnetite, partially altered to hematite, and hematite as small clusters of platy crystals.

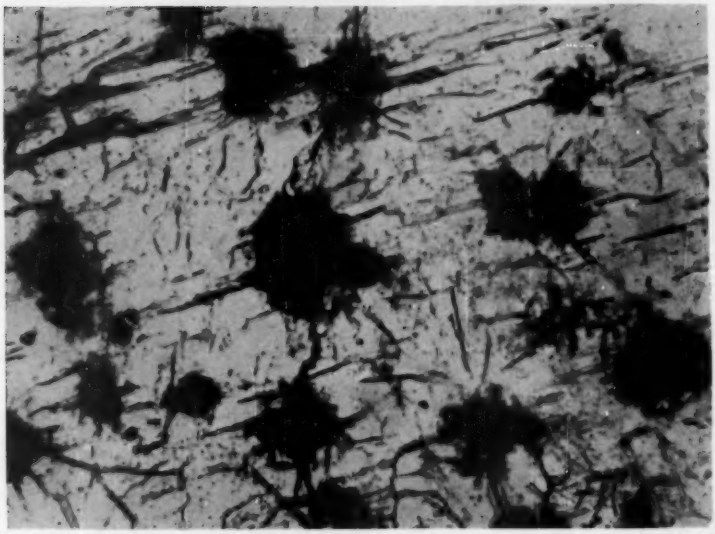
EXTINCTION HALOES

In ordinary light the albite adjacent to the uraninite grains shows little or no alteration (plate 1, A), but under crossed polars it shows prominent extinction haloes around the radioactive uraninite inclusions, as illustrated in plates 1 B and 2 A, B. These haloes have a width of 27 to 35 microns in cross-section. Within this distance from the uraninite, the extinction position of the albite differs from that of the main unaltered crystal. The difference in extinction position is greatest adjacent to the uraninite and decreases progressively outwards from it. The albite in the inner part of the halo, within about 7 microns of the uraninite, extinguishes uniformly. Infrequent orientations, in which the optic axis of an intermediate zone is nearly perpendicular to the plane of the thin section, show a range in extinction of up to 60°.

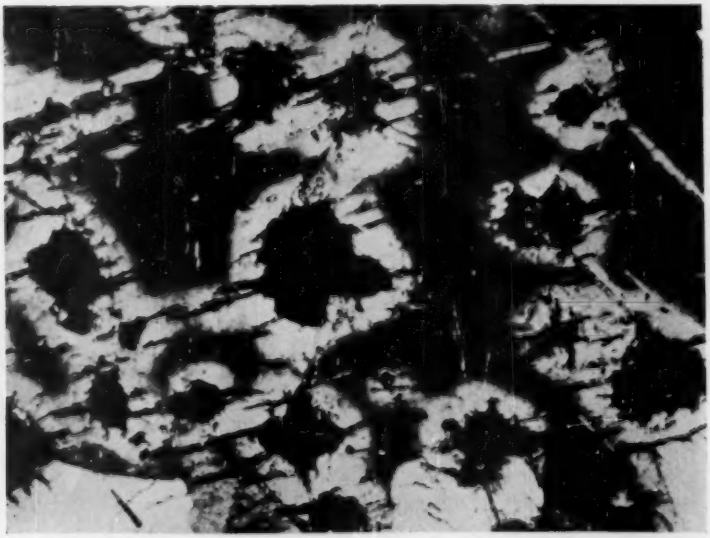
The variation of extinction angle in the halo with distance from the uraninite is shown in figure 1 for two cross-sections of different orientation. During such measurement the thin section was kept stationary, and the movement of the extinction zone determined with a micrometer eyepiece as the polars were rotated.

The curves of figure 1 show the outer limit of alteration, which in cross-sections varies between 27 to 35 microns from the uraninite. As the diameter of the uraninite is of the same order as the width of the haloes, such variation will arise from obliquity of the cross-section giving values greater than the true width. The lowest measurement of 27 microns is thus the best estimate of the width of the extinction halo. This distance is similar to or slightly less

PLATE 1

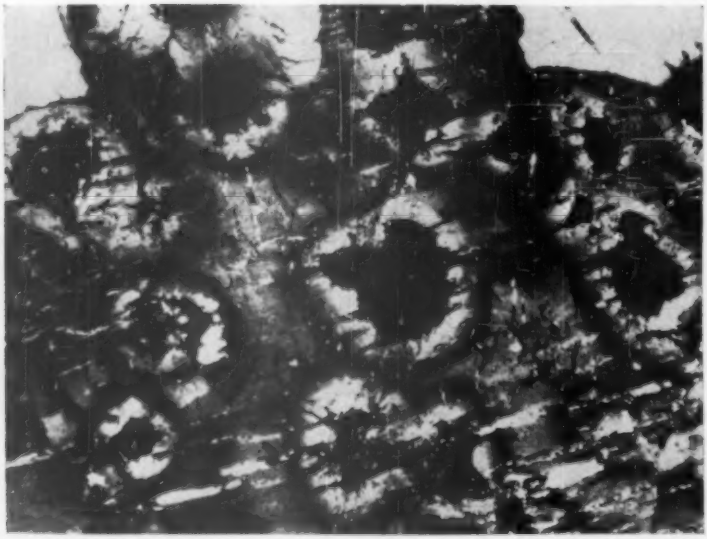


A: Uraninite inclusions (black) in albite (white). Transmitted light. ×170.

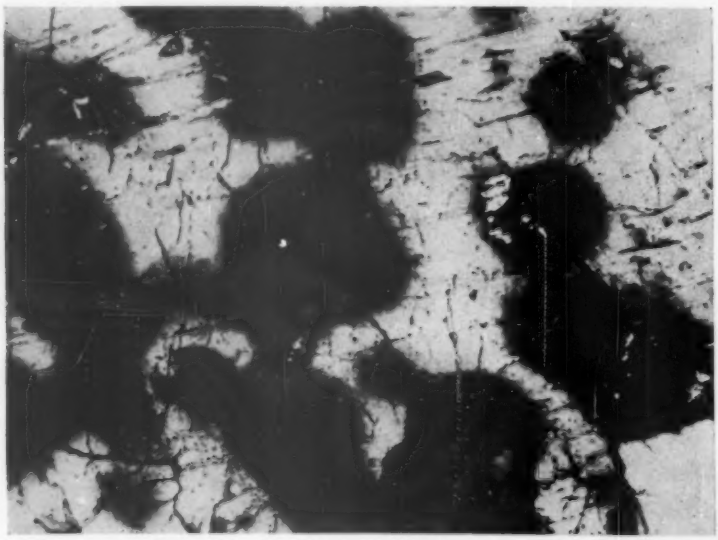


B: The same field of view as A, with crossed polars. The albite is in extinction except for the extinction haloes around the uraninite. White flecks are secondary minerals along cleavages and fractures.

PLATE 2



A: The same field of view as plate 1, with crossed polars rotated 8° from the position of plate 1 B to show the inwards displacement of the extinction zone in the halo as the polars (or the sections) are rotated.



B: The same field of view as A, with crossed polars rotated 27° from the position of plate 1 B. This rotation has displaced the extinction zone of the halo to the innermost zone adjacent to the uraninite.

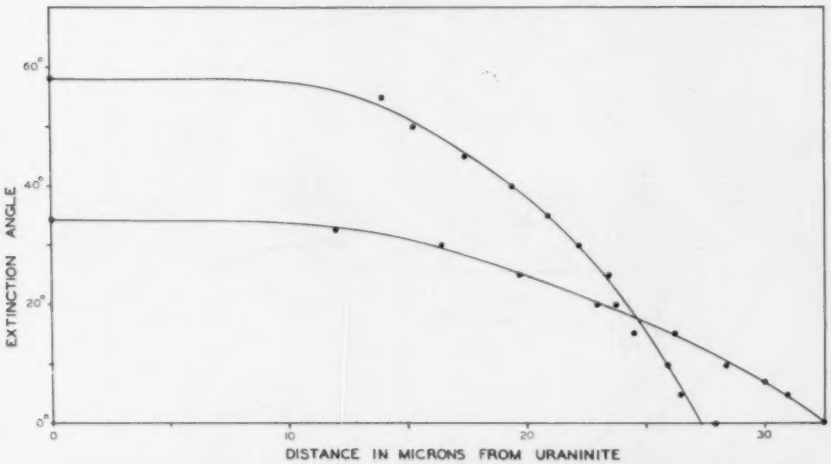


Fig. 1. Relationship between extinction angle in the halo, relative to unaltered albite, and the distance in microns of the extinction zone from the uraninite. Each curve is for a particular cross-section.

than the expected range in *albite* of the most energetic α-particles from the uranium radioactive series. The haloes are thus undoubtly produced by α-particle bombardment of the albite. Such bombardment is known as the cause of radiation damage in other crystal structures.

As measured by the extinction angle, figure 1, the optical properties change with distance from the uraninite at a rate which increases with the distance from the uraninite. This behavior is similar to the increase in declaration of α-particles with the distance travelled through a solid (Marsden and Taylor, 1913). It accords with the greater amount of ionization produced by an α-particle in the latter part of its path.

The region of the halo within about 7 microns of the uraninite shows the same, maximum, change in optical properties. Evidently the radiation damage reaches saturation at about 7 microns from the uraninite, and the increase in intensity and energy of the α-particles within this inner region is not sufficient to produce further damage. It may be noted that the decrease in velocity of the α-particles in the first 7 microns of their path through albite would be relatively low. Interpolation from the measurements of Marsden and Taylor (1913) on biotite indicates a velocity decrease of no more than 6 percent over this distance.

MOVEMENT OF OPTICAL DIRECTIONS

The movements of the optic axes and the principal optical directions associated with the change in extinction in the haloes were measured on a Leitz four axis universal stage.

Two sets of haloes were selected for measurement in which the haloes around at least two uraninite crystals lay completely within a broad albite twin lamella. The plane of (010) was determined from both the composition

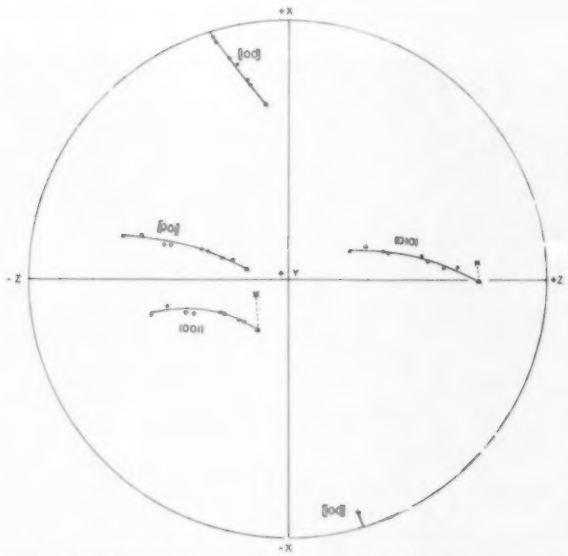


Fig. 2. Federov-Nikitin type stereogram showing the movement of the crystallographic directions in the extinction haloes relative to the principal optical directions. The unaltered, low-temperature albite is shown as a triangle, and the position of high-temperature albite by a cross.

plane and twin axis of the albite twinning in unaltered albite adjacent to the haloes; (001) was found, with somewhat less accuracy, as the mean position of a number of the cleavage planes. As the cleavages and the composition plane of the twinning pass through the extinction haloes without visible deviation their positions in the halo are essentially the same as in the adjacent, unaltered albite. The optical directions were obtained from orthoscopic settings. Careful location of an extinction ring of the halo against micro-structures in the albite was sufficiently reproducable to determine, with two exceptions, principal optical directions which were within 3° of a right angle. The measurements from two sets of haloes differ by less than the experimental error.

The optical directions of unaltered albite, and their progressive change of orientation in the extinction haloes, are shown in the stereographic projections of figures 2 and 3. The unaltered albite has the optics of low-temperature albite as given by Kaaden (1951). Radiation damage produces a continuous change in position of the principal optical directions X, Y, Z, in which X moves about 16°, while Y and Z each move about 48° from their position in low albite. The direction in the XZ plane at 20° from X to –Z moves only slightly (points A in fig. 3). The movement of the principal optical directions is thus substantially a rotation about this direction, which remains close to [4·3 $\bar{1}$ 2·2] in the crystal, with approximate coordinates $\phi = 111°$, $\rho = 88°$ for the orientation of figure 3. Such movement is different from the change to high-temperature albite, shown as a cross in figure 2.

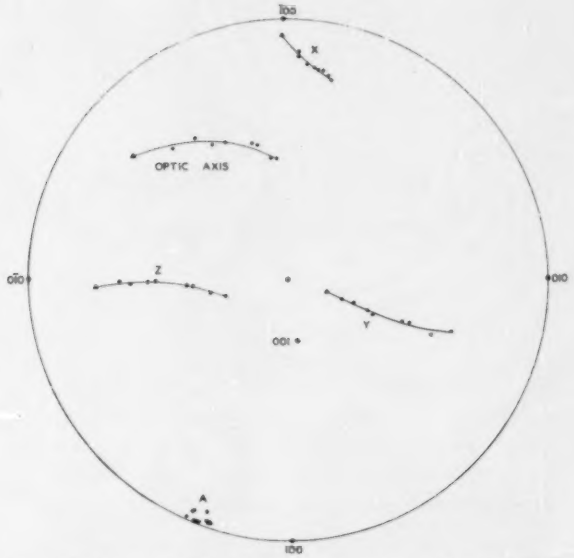


Fig. 3. Stereographic projection showing migration of the optical directions in the extinction haloes from the positions for unaltered albite (triangle). The points clustered at A are the direction in the XZ plane at 20° from X towards –Z.

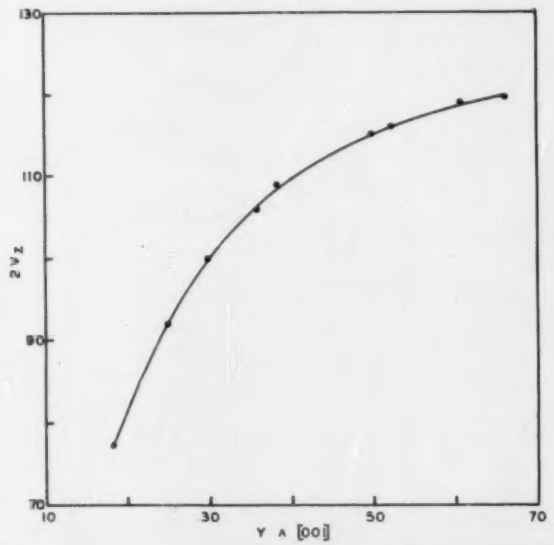


Fig. 4. Relationship between change in $2V_z$ and the rotation of the optical indicatrix, as measured by the angle from Y to [001], in the extinction haloes.

TABLE 1
Optical Directions of Unaltered and Altered Albite

Optical direction	Unaltered albite		Most altered albite	
	ϕ	ρ	ϕ	ρ
X	-90°	86°	-77°	77°
Y	15°	18°	18°	66°
-Z	178°	73°	167°	28°

The coordinates ϕ and ρ are given for the orientation of figure 3 with c vertical and ϕ measured positive clockwise from 010.

The optic axial angle changes simultaneously with the rotation of the principal optical directions, $2V_z$ increasing from 77° for the unaltered albite to 119° in the inner zones of the extinction halo. The relationship between the change of optic axial angle and the rotation of the principal optical directions as measured by the angle from Y to [001] is shown in figure 4. In the outer portion of the haloes 2V changes more rapidly than the angular rotation of the principal optical directions; with decreasing distance from the uraninite it approaches the saturation value more rapidly than does the angular rotation.

THE NATURE OF THE RADIATION DAMAGE

The optical variations associated with the radiation damage of albite lead to three main conclusions.

1. The similarity of the depth of damage to the range of the most energetic α-particles from the uranium radioactive series reveals that the damage is produced by the α-particle radiation.

2. The essential atomic structure of the albite persists in the damaged areas, as the birefringence seen in thin section does not change and saturation occurs close to the uraninite. This behavior differs from the change towards the metamict state which arises from radioactive atoms in the crystal structure, and may proceed to complete disruption of the structure.

3. The change in optical orientation produced by the α-particles is distinctly different from the change of low-temperature albite to high-temperature albite or analbite. Such difference is in accordance with the difference in structural changes normally produced in disordering developed by increased temperature, and by radiation damage.

The transformation of albite from the low-temperature to the high-temperature modifications is a change towards the equilibrium state at the higher temperature. It has been considered to arise from disordering of Si and Al in the positions of tetrahedral coordination with oxygen in the structure (Laves, 1952).

However the changes involved in radiation damage are normally towards disequilibrium. The albite of the extinction haloes has been subjected to bombardment of light charged particles, which would dissipate most of their energy by excitation of orbital electrons, with some associated removal of electrons from the outer orbits (Seitz and Koehler, 1956). A small portion of their energy is lost in direct removal of atoms from structural positions to

interstices. In addition, the removal of electrons from oxygen may disrupt the oxygen-metal bonding, and cause further movement of the oxygen to an interstitial position.

The atomic structure of albite (Taylor, Darbyshire and Strunz, 1934) contains interstitial positions adjacent to the (Si, Al)–O_4 tetrahedra which could accommodate an oxygen atom displaced from the tetrahedra. The radiation damage of albite, strikingly evident as the extinction haloes, is likely to be associated with such dislocation of oxygen, or with rearrangement of the large cation, rather than with disorder of Si and Al within structural positions.

Rotation of the optical indicatrix by radiation damage can only occur in monoclinic and triclinic minerals. The rotation in the triclinic albite is across the (010) plane, which is a plane of symmetry in monoclinic felspars. In feldspars rotation of the type here described in albite is therefore restricted to those with triclinic structures.

ACKNOWLEDGMENTS

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MINERAGRAPHIC INVESTIGATIONS
COMMONWEALTH SCIENTIFIC AND INDUSTRIAL RESEARCH ORGANIZATION
UNIVERSITY OF MELBOURNE
MELBOURNE, N.3.
AUSTRALIA

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PROPERTIES OF WATER. PART I

PRESSURE—VOLUME—TEMPERATURE RELATIONS IN STEAM TO 1000°C AND 100 BARS PRESSURE*

GEORGE C. KENNEDY

ABSTRACT. Previous data of the author on the P-V-T relations of water were seriously in error in the high temperature-low pressure field. The error, caused by reaction between the water and the wall of the experimental bomb, has now been corrected by adding an oxidizing agent (Cu O) to the charge, and corrected data can now be presented.

INTRODUCTION

Growing use of thermodynamic theory in the solution of geological and mineralogical problems (for example, MacDonald, 1955; Majumdar and Roy, 1957) as well as improved experimental techniques has made necessary precise values for the density of various geologically important gases at high pressures and temperatures.

Steam is the most abundant natural gas, and accurate measurements of pressure-volume-temperature relations in steam are of fundamental importance. Keyes Smith and Gerry (1936) studied the P-V-T relations of steam to 360 bars and 460°C. Kirillin and Rumyantsev (1950) and Kirillin, Rumyantsev and Zubarev (1956), extended the temperature range to 650°C, and Kennedy (1950) extended it further to 1,000°C. Recent determination of specific volume of water at high temperatures and pressure (Holser and Kennedy, 1956; Kirillin, Rumyantsev and Zubarev, 1956), suggest that the Kennedy (1950) water data is accurate to within one-half to one percent or less in the pressure region 300-1400 bars at high temperatures, but that the data are much in error for pressures below 300 bars at high temperatures. This error reaches a maximum of about 10 percent in the volume term at 1,000°C and 100 bars. Such a large error could have readily been detected if a plot of PV/RT had been made for the 1950 data; unfortunately, however, no such plot was made. The purpose of this paper is to present the results of some new measurements of P-V-T relations in steam to 100 bars pressure and 1,000°C, using methods modified to avoid the aberrant results at the lower pressures reported in 1950.

EXPERIMENTAL EQUIPMENT

The furnaces, bombs and methods of temperature regulation and measurement used in this investigation have all been described elsewhere (Kennedy, 1954). The only departure from the equipment assemblage described in 1954 was the addition of a dead-weight piston gage to the 100-bar Heise gage used in the earlier pressure measurements, so that regular intercomparisons could be made with the Heise gage for close control of drift and calibration.

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SOURCE OF ERROR IN EARLIER WORK

The errors in the observations reported in 1950 seem to have been due largely to interaction between water and the bomb alloy, with the resultant production of hydrogen and metal oxides. Water extracted from the bomb, after prolonged contact with the bomb at 800° to 1,000°C, contained abundant hydrogen, so that the water effervesced violently on cooling and release of pressure. This resulted in computation of spuriously large fluid densities from the volumeter readings at low pressure as the hydrogen exsolved. The reaction is presumed to be as follows:

$$4H_2O + 3R = 4H_2 + RO + R_2O_3$$

Where R may be Fe, Cr, Ni, and Co of the bomb alloy.

CORRECTION OF ERROR

The problem of interaction between water and the bomb walls may be avoided in at least two ways. The bomb may be protected by a heavy lining of platinum or gold that will not oxidize appreciably, or the reaction may be prevented or counterbalanced by keeping the water oxidized. Attempts to prevent the reaction by lining the bomb with platinum or other inert metal were not entirely successful. Establishing a gas-tight seal between a platinum liner and the bomb closures is most difficult, as is the protection of thermocouple tubes and pressure inlets from contact with water. Further, unless the liner is perfectly fitted to the bomb, gas is likely to be trapped behind the liner forcing it to act as a bellows as pressure and temperature changed. This resulted in uncertainty as to the volume being measured.

The second alternative, to keep the water oxidized, involves no mechanical complications. It is based on the introduction of a few grams of a finely divided oxidizing agent, such as CuO, into the bomb chamber along with the water. At high temperatures, CuO reacts with hydrogen released by the oxidation of the bomb walls:

$$H_2O + R \rightarrow RO + H_2$$
$$H_2 + CuO \rightarrow H_2O + Cu$$

In this system, CuO is converted to Cu at essentially the same rate as the metals of the bomb alloy are oxidized and the volume of H_2O remains unchanged. Water, extracted from the bomb after prolonged contact with the bomb alloy at high temperatures, shows no appreciable hydrogen.

METHOD

In the present investigation, runs were made following experimental methods described in detail previously (Kennedy, 1954) but with the addition of a few grams of finely-divided CuO to the bomb load. In brief, the method is as follows: An unknown quantity of water is injected into a bomb of unknown size and a series of simultaneous observations of pressure and temperature made. This approximate isochore is then plotted on large-scale graph paper and graphical corrections for pressure and temperature dilation of the bomb made. The resulting plot is of an iso-volumetric curve for varying T and P. However, the specific volume for the particular iso-volumetric plot is unidentified. The highly precise data of Keenan and Keyes (1936) for lower

temperature, is then used to identify the iso-volumetric curves as to specific volume. This is done by taking from the graphical isometric plot the precise pressure at which the isochore crosses the 610° isotherm of Keenan and Keyes (approximately 312°C). Reference to the data of Keenan and Keyes then gives the specific volume for the combination of pressure and temperature.

To check the accuracy of measurements made with the bomb protected by CuO, a series of simultaneous observations of pressure and temperature were made at successively higher temperatures with a fixed quantity of water and then at a series of successively lower temperatures with the same quantity of water. The observations on the rising temperature runs fell along a smooth curve. On the return series of observations, made after the bomb and its contained water charge were allowed to remain at high temperature for some 40 hours, the set of observations fell along the same smooth curve. Similar runs in unprotected bombs gave lower pressures on the return set of observations. In addition, the dew point in unprotected bombs was much obscured.

Various additional evidence, internal to the body of data, indicates that errors arising from reaction between water and wall are eliminated, or nearly so, by the use of CuO. For example, the value PV/RT computed from old high-temperature data from unprotected bombs does not extrapolate to 1.000

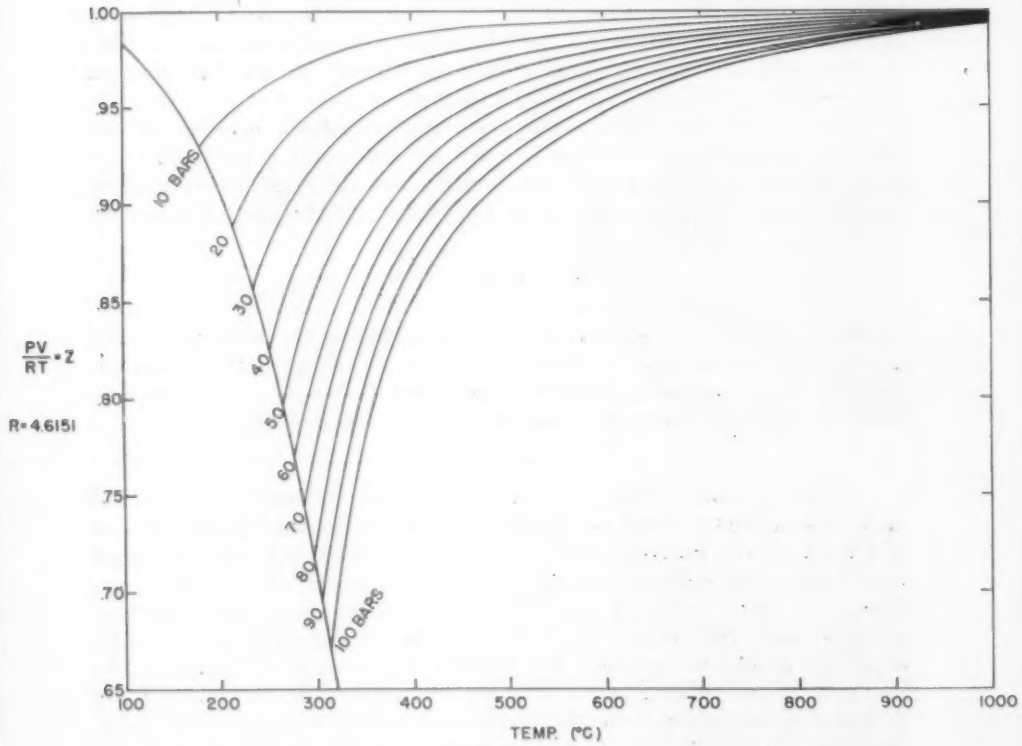

Fig. 1

for zero pressure, but in protected bombs, values at various isotherms of PV/RT trend smoothly toward 1.000 for successively lower pressures.

RESULTS

A large number of curves of T versus P were determined experimentally for specific volumes ranging from 20 to 500 cc/gm. These curves are essentially straight lines giving the best fit to the series of observations of P, at various T values, for a given quantity of steam in the bomb. Values of PV/RT were then computed for various isotherms from these data and adjusted numerically to yield smooth tables of differences along the various isotherms and isobars. The data of Osborne, Stimson and Ginnings (1939) were used as a guide along the saturation curve. Specific volumes were then computed from the adjusted values of PV/RT and are entered in table 1. Deviation from ideal gas behavior is shown in figure 1.

COMPARISON OF RESULTS WITH OTHER DATA

The tables herein presented overlap the data of Keyes and associates to 460° and also may be compared with their extrapolated data. In addition, they may be compared with the few measurements for this pressure range to 650°C by Kirillin and associates. The Keyes' data have been converted to the metric system by Faxen (1953) and his tabulation is used here for comparison (table 2). The Kirillin data is recorded in units and intervals which make direct comparison difficult. To ease comparison, matching values have been graphically interpolated on the curves presented by Kirillin.

The data, measured and extrapolated, of Keyes agree very well with the smoothed values presented here. The agreement is to 1/1000 or better almost everywhere in the range where measurements overlap. This is not surprising as the new data are tied to the 312°C isotherm of Keyes. However, the close agreement of the present data with Keyes' extrapolation above his highest measurement at 460°C encourages confidence in both Keyes' extrapolation and the new results.

The Kirillin values and the present measurements are in excellent agreement at 400° and 450°, but at higher temperatures the departure between the two sets of data is considerable. Some recent measurements by Holser and Kennedy (1956) at higher pressures are, however, in good agreement with the Kirillin higher pressure values, so that the major disagreement with Kirillin's data is restricted to the low pressure—high temperature region. This suggests that Kirillin may have encountered, in some measure, the problem of reaction between water and bomb wall.

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TABLE 1

Specific Volume of Steam/Pressure in Bars

Temp. °C	10	20	30	40	50	60	70	80	90	100
200	205.8									
210	211.4									
220	216.9	102.25								
230	222.3	105.42								
240	222.7	108.47	68.18							
250	232.9	111.44	70.58							
260	238.9	114.35	72.85	51.80						
270	243.9	117.20	75.03	53.69	40.60					
280	249.5	119.99	77.14	55.50	42.27	33.18				
290	252.9	122.74	79.18	57.22	43.85	34.74	28.04			
300	257.9	125.45	81.15	58.86	45.34	36.17	29.47	24.25		
310	262.9	128.13	83.08	60.44	46.76	37.50	30.76	25.59	21.42	
320	267.8	130.78	84.98	61.98	48.12	38.75	31.97	26.80	22.67	19.26
330	272.7	133.41	86.85	63.49	49.43	39.95	33.11	27.92	23.81	20.44
340	277.6	136.02	88.72	64.97	50.71	41.11	34.20	28.97	24.85	21.49
350	282.4	138.61	90.55	66.43	51.96	42.23	35.25	29.96	25.80	22.44
360	287.2	141.18	92.36	67.88	53.18	43.31	36.25	30.91	26.70	23.32
370	292.1	143.74	94.13	69.31	54.37	44.37	37.21	31.82	27.57	24.15
380	297.0	146.20	95.86	70.71	55.54	45.41	38.15	32.69	28.40	24.94
390	301.8	148.74	97.63	72.09	56.69	46.43	39.07	33.53	29.19	25.69
400	306.6	151.2	99.38	73.45	57.83	47.43	39.96	34.33	29.95	26.41
410	311.4	153.7	101.11	74.79	58.95	48.42	40.83	35.12	30.68	27.11
420	316.2	156.2	102.83	76.12	60.05	49.38	41.68	35.90	31.40	27.79
430	321.0	158.7	104.53	77.44	61.13	50.32	42.51	36.67	32.11	28.46
440	325.8	161.2	106.32	78.75	62.20	51.24	43.33	37.42	32.81	29.11
450	330.6	163.6	108.00	80.04	63.27	52.13	44.15	38.16	33.50	29.75
460	335.4	166.1	109.67	81.33	64.33	53.03	44.96	38.90	34.18	30.38
470	340.2	168.5	111.33	82.61	65.39	53.90	45.76	39.62	34.85	31.00
480	344.9	170.9	112.98	83.89	66.44	54.81	46.55	40.33	35.51	31.61
490	349.6	173.3	114.52	85.16	67.49	55.72	47.33	41.03	36.16	32.21
500	354.3	175.7	116.15	86.42	68.53	56.63	48.10	41.74	36.79	32.79
510	359.0	178.1	117.88	87.67	69.56	57.51	48.86	42.43	37.41	33.36
520	363.7	180.5	119.50	88.92	70.58	58.39	49.62	43.11	38.02	33.92
530	368.4	182.9	121.12	90.16	71.60	59.36	50.44	43.78	38.62	34.48
540	373.1	185.2	122.73	91.41	72.62	60.12	51.17	44.45	39.22	35.03
550	377.8	187.7	124.34	92.65	73.63	60.97	51.91	45.11	39.81	35.58
560	382.5	190.1	125.95	93.89	74.63	61.81	52.64	45.76	40.40	36.13

TABLE 1 (Continued)

Specific Volume of Steam/Pressure in Bars

Temp. °C	10	20	30	40	50	60	70	80	90	100
570	387.2	192.5	127.57	95.12	75.63	62.65	53.37	46.42	40.99	36.68
580	391.8	194.8	129.15	96.34	76.63	63.49	54.11	47.08	41.58	37.23
590	396.5	197.2	130.73	97.56	77.63	64.32	54.84	47.73	42.17	37.77
600	401.2	199.6	132.31	98.77	78.62	65.15	55.57	48.38	42.76	38.31
610	405.8	201.9	133.89	99.98	79.61	65.99	56.30	49.03	43.35	38.85
620	410.4	204.2	135.48	101.19	80.60	66.83	57.03	49.68	43.94	39.38
630	415.0	206.6	137.1	102.40	81.59	67.66	57.76	50.33	44.52	39.91
640	419.7	209.0	138.7	103.60	82.58	68.50	58.48	50.97	45.10	40.44
650	424.3	211.3	140.3	104.81	83.56	69.33	59.20	51.61	45.68	40.96
660	428.9	213.6	141.9	106.02	84.54	70.16	59.92	52.25	46.36	41.48
670	433.6	216.0	143.5	107.23	85.52	70.99	60.64	52.88	46.83	42.02
680	438.3	218.4	145.1	108.43	86.50	71.81	61.35	53.51	47.40	42.53
690	443.0	220.7	146.7	109.63	87.47	72.62	62.06	54.14	47.97	43.04
700	447.6	223.1	148.3	110.83	88.44	73.43	62.77	54.76	48.54	43.55
710	452.2	225.4	149.9	112.03	89.41	74.25	63.47	55.38	49.10	44.06
720	456.9	227.8	151.5	113.23	90.38	75.06	64.17	56.00	49.65	44.56
730	461.5	230.1	153.0	114.43	91.35	75.87	64.87	56.62	50.20	45.06
740	466.2	232.5	154.5	115.63	92.32	76.68	65.57	57.25	50.76	45.56
750	470.9	234.8	156.2	116.83	93.29	77.49	66.27	57.87	51.31	46.06
760	475.6	237.2	157.9	118.03	94.26	78.30	66.97	58.48	51.86	46.56
770	480.2	239.5	159.4	119.22	95.22	79.11	67.67	59.09	52.41	47.06
780	484.8	241.9	160.9	120.41	96.18	79.92	68.36	59.70	52.96	47.56
790	489.4	244.2	162.5	121.60	97.14	80.73	69.06	60.31	53.51	48.06
800	494.1	246.6	164.1	122.79	98.10	81.54	69.76	60.92	54.06	48.55
810	498.8	248.9	165.7	123.98	99.06	82.35	70.46	61.53	54.51	49.04
820	503.5	251.3	167.2	125.17	100.02	83.16	71.16	62.14	55.15	49.54
830	508.1	253.6	168.8	126.36	100.98	83.96	71.86	62.75	55.70	50.03
840	512.8	255.9	170.3	127.55	101.94	84.77	72.55	63.35	56.24	50.52
850	517.5	258.3	171.9	128.74	102.90	85.57	73.24	63.97	56.79	51.01
860	522.2	260.6	173.5	129.93	103.85	86.38	73.93	64.58	57.33	51.50
870	526.8	263.0	175.0	131.12	104.80	87.18	74.62	65.19	57.87	51.99
880	531.4	265.3	176.6	132.31	105.75	87.98	75.31	65.79	58.41	52.47
890	536.0	267.7	178.2	133.50	106.70	88.78	76.00	66.40	58.95	52.96
900	540.6	270.0	179.8	134.68	107.65	89.58	76.69	67.00	59.49	53.47
910	545.2	272.4	181.3	135.87	108.60	90.38	77.38	67.61	60.03	53.95
920	550.1	274.8	183.9	137.06	109.55	91.18	78.07	68.21	60.57	54.46
930	554.7	277.1	184.5	138.24	110.50	91.98	78.76	68.82	61.11	54.94
940	559.3	279.4	186.0	139.42	111.45	92.78	79.44	69.43	61.65	55.43
950	563.9	281.7	187.6	140.61	112.40	93.58	80.12	70.03	62.19	55.91
960	568.5	284.1	189.2	141.79	113.35	94.38	80.80	70.64	62.73	56.40
970	573.2	286.4	190.7	142.98	114.30	95.18	81.48	71.25	63.27	56.89
980	578.1	288.8	192.3	144.16	115.25	95.97	82.16	71.86	63.81	57.40
990	582.7	291.2	193.9	145.35	116.20	96.76	82.86	72.47	64.35	57.89
1000	587.3	293.5	195.5	146.53	117.15	97.55	83.56	73.08	64.89	58.38

TABLE 2

Intercomparison of Recent Steam Data

		Kennedy (1957)	Keyes et al (1936)		Kirillin et al (1956)
	50 bars	57.83	57.81		—
400°C					
	100 bars	26.41	26.42		26.42
	50 bars	63.27	63.29		—
450°C					
	100 bars	29.75	29.75		29.77
	50 bars	68.53	68.53	Extrapolated Values	—
500°C					
	100 bars	32.79	32.76		32.85
	50 bars	73.63	73.63		—
550°C					
	100 bars	35.58	35.59		35.69
	50 bars	78.62	78.62		—
600°C					
	100 bars	38.31	38.30		38.51
	50 bars	83.56	83.54		—
650°C					
	100 bars	40.96	40.93		39.97

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INSTITUTE OF GEOPHYSICS
UNIVERSITY OF CALIFORNIA
LOS ANGELES, CALIF.

In Memoriam

REGINALD A. DALY
1871-1957

Professor Reginald A. Daly, the senior Associate Editor of this JOURNAL, died on September 14, in Cambridge, Mass., at the age of 86. He was one of the great, and by the same token controversial, figures in geology in the first half of the century. Few men have contributed ideas to more diverse fields of geology, ideas that for the most part met not acceptance but vigorous dissent. Piecemeal magmatic stoping, glacial control of coral reefs, density currents in submarine canyons, compositional shells in the interior of the Earth—major controversies have raged over all these suggestions, yet even antagonists must concede the fruitfulness of Daly's ideas, not merely in stirring up others to study the phenomena in question, but also in contributing to their ultimate explanation, which has generally turned out to be more complex than either Daly or his opponents thought.

He has been stimulating also to the AMERICAN JOURNAL OF SCIENCE, and he was one of our last links with the days of the Danas, when the JOURNAL was still a private venture. We are proud to have published his articles over a span of almost fifty years, including some of the most famous and controversial. We are proud to have published in 1945 the "Daly Volume", dedicated to him on the occasion of his retirement. What Daly contributed to geology is best revealed perhaps in the series of excerpts from his writings, entitled "Strategy in Geology", that introduced that volume—excerpts that reveal the combination of controlled hypothesis, speculation, and imagination that made him so stimulating a leader of our science.

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The Centre National de la Recherche Scientifique de France has announced the publication of a very large part of the International Lexicon of Stratigraphy, which is being prepared by collaborators all over the world, under the guidance of a French committee and the energetic direction of Dr. Jean Roger of the Centre d'Etudes et Documentation Paléontologiques. Fascicles for the various countries are being published as they are ready; they are mainly in the language of the country or French or English; the lexicon for the U.S.S.R. will appear in French translation. Fascicles can be bought individually, or orders can be placed for whole volumes at the following prices:

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